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SYNOPTIC TO MESOSCALE DYNAMICS
OF SEVERE THUNDERSTORM ENVIRONMENTS:
A DIAGNOSTIC STUDY WITH FORECASTING APPLICATIONS

by
G.S. STRONG

A THESIS
SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH
IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE OF

DOCTOR OF PHILOSOPHY
in
METEOROLOGY

DEPARTMENT OF GEOGRAPHY

Edmonton, Alberta

Spring, 1986

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The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled SYNOPTIC TO MESOSCALE DYNAMICS OF SEVERE THUNDERSTORM ENVIRONMENTS: A DIAGNOSTIC STUDY WITH FORECASTING APPLICATIONS, submitted by GEOFFREY S. STRONG in partial fulfilment of the requirements for the degree of DOCTOR OF PHILOSOPHY in Meteorology.

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DEDICATION

Dedicated with love to my family:

to my wife, Phyllis, and my three daughters, Emma, Michelle, and Jennifer, without whose constant support, love, and patience, I could not have completed this work; also to my parents, who early in my life, taught me the value of education and stubborn persistence.

ABSTRACT

The goal of this research is to investigate the influences of synoptic scale dynamics on severe mesoscale thunderstorm systems. An extensive literature review is undertaken, which provides the convective storm forecasting background. The review includes climatological and kinematic studies, and case studies using the author's own forecasting technique. This is followed by a theoretical review of the interactions among various scales of atmospheric motion, including some diagnostic results relevant to interactions between the synoptic scale and the mesoscale.

This review provides a basis upon which to build the hypothesis that synoptic influences extend beyond pre-storm organization and convective triggering of the mesoscale moisture field, to also include a significant control over severe storms throughout their life cycle. The concept of a downscale transfer of energy to the mesoscale is also embodied in the hypothesis, which focuses on the creation and breakdown of the capping inversion and its role in the formation and life cycle of a severe storm complex.

The hypothesis is tested in three stages. First, using several Oklahoma-Texas severe storm cases, low-level ascent and surface convergence are shown to exist over mesoscale regions at least six hours prior to storm formation. It is shown that these mesoscale circulations are due in part to upstream synoptic scale processes, and that the ascent weakens the capping lid. Second, various thermodynamic

instability, moisture, and wind fields are considered in relation to the computed vertical motion fields. The relative merits of using these fields for storm prediction are discussed. Third, the author's technique for predicting Alberta convective storms, based primarily on synoptic scale variables from a different data source, is successfully tested on the Oklahoma-Texas storms.

The qualitative agreement between these various analyses, the forecast results, and the observed synoptic and convective cloud systems, provides substantial evidence for the postulated synoptic influences. A further extension of this conceptual model relates the decay and growth stages of the various scales of motion to preferences for downscale or upscale interactions with adjacent scales. It suggests how this might contribute to severe convective storms. Finally, recommendations are made for future field research on mesoscale convective systems, and for additional tests of the hypothesis.

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The impetus for this study came from twenty years experience in forecasting and related work, first with the Atmospheric Environment Service of Canada, and since 1976, with the Alberta Research Council. I wish to acknowledge both of these organizations, especially ARC, which recognized the value of mesoscale research to its own goals, and provided the necessary financial assistance and computing resources while I continued full-time employment.

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ABBREVIATIONS used in Text

AC	- altocumulus clouds
ACC	- altocumulus castellanus clouds
ACSL	- altocumulus standing lenticularis clouds
AES	- Atmospheric Environment Service (of Environment Canada)
AHP	- ARC Hail Project, Red Deer, Alberta
AKE	- available kinetic energy
APE	- available potential energy
ARC	- Alberta Research Council, Edmonton, Alberta
AVE	- Area Variability Experiment
C	- degrees Celsius
CB	- cumulonimbus cloud
CCOPE	- Cooperative Convective Precipitation Experiment
CDC	- Convective Day Category
CIS	- Capping Inversion Study, 1982 (of AHP)
CISK	- Conditional Instability of the Second Kind
CMC	- Canadian Meteorological Centre (of AES)
CS	- cirrostratus clouds
CU	- cumulus clouds
g kg ⁻¹	- grams/kilogram
GMT	- Greenwich Mean Time, now known as UTC
GOES	- Geostationary Operational Environment Satellite
K	- degrees Absolute
KE	- kinetic energy
KT75	- Kung and Tsui (1975)
LIMEX	- Limestone Mountain Experiment (of AHP)
LJ	- lower jet (or streak)
mb	- millibar(s)
MCC	- Mesoscale Convective Complex
MK72	- McInnis and Kung (1972)
m s ⁻¹	- meters/second
NCAR	- National Center for Atmospheric Research, Boulder, Colorado
NHRE	- National Hail Research Experiment of NCAR
NOAA	- National Oceanic and Atmospheric Administration
NSSL	- National Severe Storms Laboratory, Norman, Oklahoma
NVA	- negative vorticity advection
NWS	- National Weather Service (of NOAA)
PVA	- positive vorticity advection
RW	- (moderate) showers (with qualifiers -, +, ++ for light, heavy, or very heavy)
SESAME	- Severe Environmental Storms and Mesoscale Experiment
SC	- stratocumulus cloud
Sc3	- Synoptic Index of Convection (3-predictor)
Sc4	- Synoptic Index of Convection (4-predictor)
Sc10	- Synoptic Index of Convection (10-predictor)
SSCC	- Sub-synoptic Scale Convective Complex
TCU	- towering cumulus, or cumulus congestus clouds
TK77	- Tsui and Kung (1977)
Theta	- potential temperature
Theta-E	- equivalent potential temperature
Theta-W	- Pseudo-adiabatic wet-bulb potential temperature

TRW - (moderate) thundershower (with qualifiers -,+,++ for light,
heavy, and very heavy (or severe)
UJ - upper jet (or streak)
UTC - Coordinated Universal Time, formerly known as GMT or Z
W - watts
Z - abbreviation on maps for UTC

"When one considers the fact that the upper winds steadily blow through the troughs aloft, one is forced to conclude that the planetary wave disturbances of middle latitudes cannot be considered as closed systems, but that they transmit energy to, or receive it from, the surrounding atmosphere".

C.-G. Rossby (1949)

CHAPTER 1.

INTRODUCTION

1.1 THUNDERSTORMS AND THE CONCEPT OF SCALE INTERACTIONS

More than 35 years ago, it was shown that the planetary and synoptic scale processes interact through exchanges of potential and kinetic energy (Starr, 1948; Rossby, 1949). Diagnostic studies over the past 30 years have confirmed earlier theories in this regard, and have led some scientists, even as early as White and Saltzman (1956), to speculate on possible interactions between the synoptic scale and severe extra-tropical thunderstorm systems. These "scale interactions", as they have come to be known, include both the downscale and feedback modes. For example, Sechrist and Dutton (1970) showed that a developing shortwave cyclone can extract upper level kinetic energy from the mean planetary flow upstream (the downscale process), while converting potential energy at low-levels in the atmosphere to kinetic energy of the mean flow at intermediate levels downstream (the upscale process).

Apart from several exploratory studies by Kung and his associates at the University of Missouri during the 1970s, diagnostic studies of energetics at the smaller scales have been limited to investigations of energy "feedback" from the severe storm to the mean flow (e.g., Fuelberg and Printy, 1983, 1984). Since frictional dissipation alone cannot account for all of the released latent heat once the storm dies, a feedback of energy to the synoptic scale flow seems reasonable. However, this does not complete the conceptual model by any means. "Downscale" energy transfers, from synoptic processes to the mesoscale, and especially to the pre-storm environment, have been largely ignored.

Barnes (1976) reviewed many of the conceptual models of thunderstorms which have evolved over the past century. To a forecaster, the most striking similarity in these models is the concentration on the visible, in-cloud processes. There appears to be general agreement that the micro-physical processes involved are essentially the same for the severe storms of Alberta, Colorado, and Oklahoma (Krauss and Marwitz, 1984). However, there is no universal agreement on the relative importance of these processes. Also, models of the kinematics of storm structures and their immediate environmental flow show marked variations over the past dozen years (e.g., Chisholm and Renick, 1972; Lemon and Doswell, 1979; Krauss and Marwitz, 1984). There is even less agreement on the degree of importance of the larger scale (synoptic) processes in convective storm development.

The goal of the current study is thus to consider several aspects of scale interactions, and how these might pertain to the evolution of convective storms. As a forecaster of some 20 years, the author is particularly interested in how such scale interactions affect the

pre-storm environment of severe thunderstorm systems. In order to begin, it is necessary to discuss the scales of atmospheric motion being considered. The literature review, commencing in Section 1.3, will then attempt to show how we have arrived at our present level of understanding of scale interactions involving severe storms, and why the research of this report was undertaken. Chapter 2 will deal with energetics theory and recent diagnostic studies, and how these provide a means to investigate the physical interactions of synoptic processes with convective storms. The rather extensive literature review is necessary background for a conceptual model of scale interactions developed in Chapter 3, and for the analyses of this thesis which follow in Chapters 4-7.

1.2 SCALES OF ATMOSPHERIC MOTION

One scheme for classifying the atmospheric 'scales' of motion is shown in Table 1.1. These scales are not discrete physical entities, but rather, they represent the most common stratifications discussed in the literature (e.g., Orlanski, 1975). The reader is advised that many somewhat different variations of this table appear in the literature.

TABLE 1.1: Definitions and examples of the scales of atmospheric motion discussed in this paper, following closely those of Orlanski (1975).

SCALE	Range of WAVELENGTH (km)	Time Scale of Associated Weather over region	Scale EXAMPLES	Example Data Source	Mean Spacing or Data Resolution (km)	Frequency of Observation	REMARKS
MACRO or PLANETARY	> 5000	More than one week	Longwaves, large, persistent quasi-stationary highs, lows	Global upper air network (weekly averages)	500 (land stations)	twice daily; 00Z, 12Z	variable density; e.g., often inadequate over oceans
SYNOPTIC (upper Meso- α)	2000 to 5000	2-7 days	Intermediate and shortwaves, migratory lows, highs	N.A. upper air network	500	twice daily; 00Z, 12Z	irregular design; e.g., no sites over Saskatchewan
SUB-SYNOPTIC (lower Meso- α)	200 to 2000	12-48 hours	Sub-synoptic Convective Complexes (SSCCs), or TRW clusters, Hurricanes, 2ndry lows	N.A. surface observing network	130	hourly plus special observations	irregular design, sites sometimes not representative
MESO (Meso- β)	20 to 200	1-12 hours	Large TRWs or Mesoscale Convective Complexes (MCCs), cold fronts and squall lines, cumulus clusters	NSL (1966-67) and SESAME (AVE-III/IV/V) upper air networks Alta. Forestry reports	75 80 30	1.5-3 hours 1.5-3 hours twice daily; 14Z, 19Z	very irregular design; near-regular design; irregular, limited area (Alberta foothills)
CLOUD (Meso- γ) or CUMULUS Scale	2 - 20	10-60 minutes	Single-cell CU, TCU, small CB, tornado, fine-scale reflectivity pats.	GOES sat. data ARC raingauges NOAA sat. data	4-16 (IR) 1-2 (VSB) 2-10 1	30 minutes 30 minutes once daily 4-8 hours	resol. detectors. N from equator AMP region Polar-orbiting
MICRO	few cm to 2 km	seconds to minutes	turbulent eddies, convective thermals, CU cloud turrits, wind gusts	Radar data LANDSAT data Research aircraft data	0.5-2.0 0.05 <0.01	1 minute 1-2 weeks 1 second	uniform grid poor temporal uniform, but 1-D unless sev. acft
VISCOUS	molecular distances	??	molecular motions	electron microscope(?)	N/A for most meteorology applications	N/A	N/A

Examples of three of these scales of motion are revealed in the enhanced infrared satellite image of Figure 1.1. It includes a synoptic scale cyclonic system of more than 2000 km across, several sub-synoptic (or Meso- α) scale convective complexes (clusters of thunderstorms) with horizontal dimensions of 250-350 km in this case, and at least one mesoscale (Meso- β) thunderstorm complex of 50-100 km. The approximate position of the surface front is superimposed on Figure 1.1. The aerial photograph of Figure 1.2 depicts three other scales of motion from a much closer vantage point. Here the thunderstorm system is a (Meso- β) mesoscale convective complex of 50-80 km across. This complex includes cloud scale (Meso- γ) clusters of 2-10 km across, which by themselves would be identified as cumulus congestus (TCU) clouds, each of which is made up of microscale cloud turrets of the order of 1 km or less across. The precipitation patterns from these clusters have been observed on radar as 'fine-scale reflectivity patterns' or FSPs (Barge and Bergwall, 1976). It has been assumed in the past that these clusters, and certainly the turrets, are a result of turbulent processes (e.g., Tennekes, 1978). Still, Barge and Bergwall have shown that the FSPs exhibit continuity in space and time, suggesting that such features have predictable qualities. This may require some re-examination of those features which we assume to be indeterminate three-dimensional turbulence, a problem which cannot be resolved in the current study.

A further clarification of the two convective complex scales is necessary here. Maddox (1980) defined but one scale of Mesoscale Convective Complex (MCC), having a quasi-elliptic shape and a total cloud shield area exceeding 100,000 km² (linear horizontal dimension

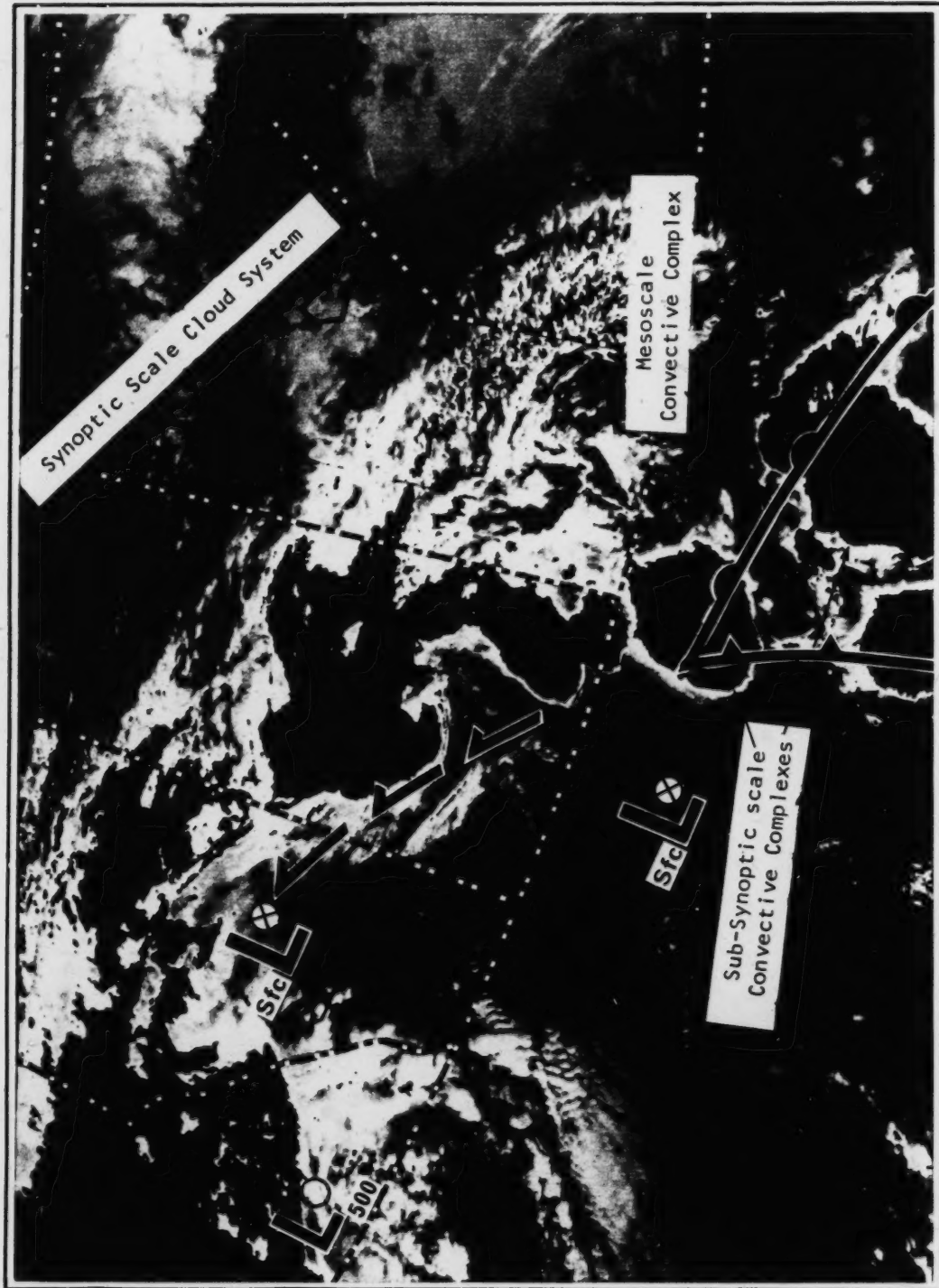


FIGURE 1.1: Enhanced infrared NOAA-6 image for 0136 UTC, 15 July, 1982, depicting three of the scales of atmospheric motion indicated in Table 1.1.

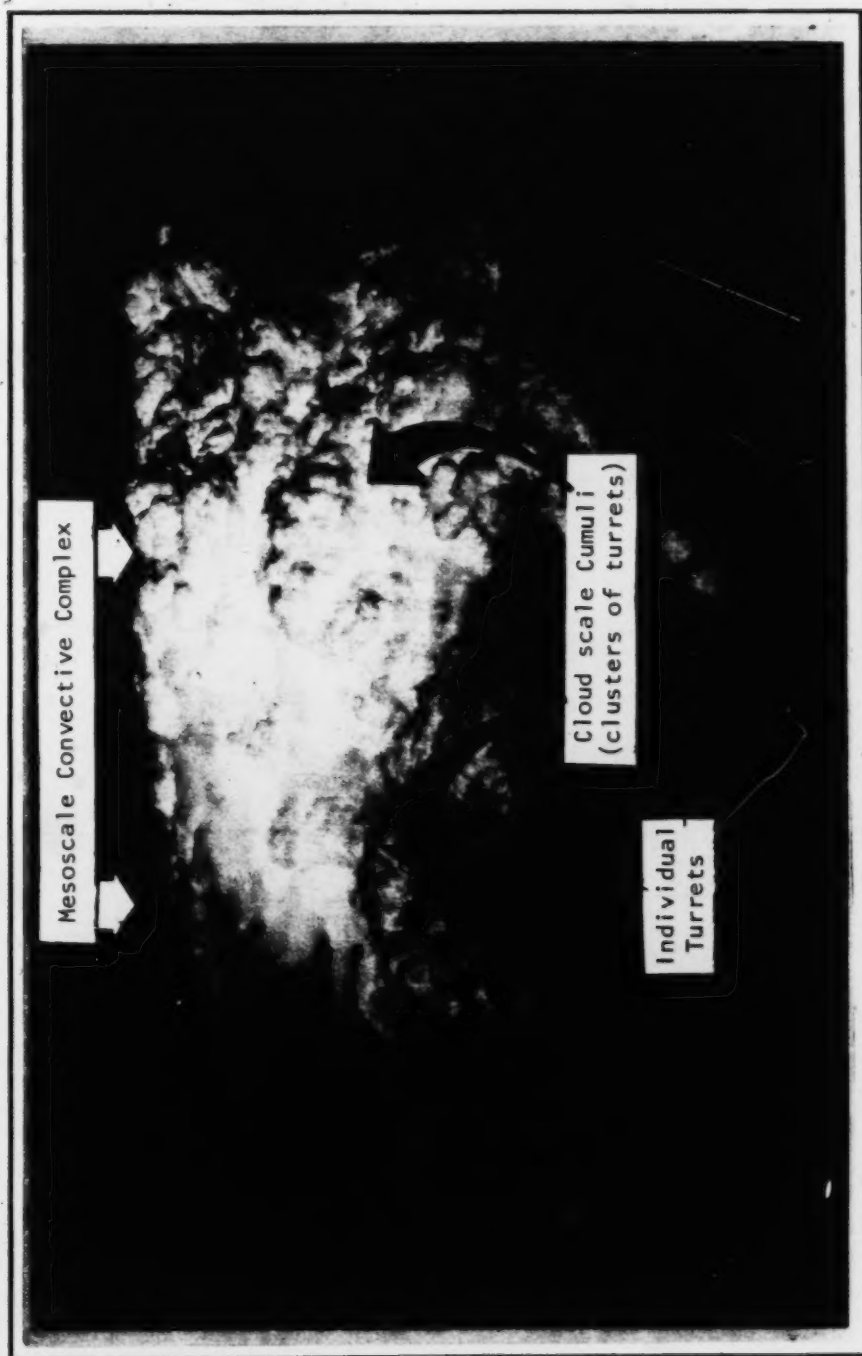


FIGURE 1.2: Aerial view of a mesoscale thunderstorm complex (40-80 km across), with cloud scale clusters (2-10 km across) in the "new growth zone", and individual cloud turrets with horizontal dimensions of 1 km or less - three identifiable interacting scales of motion.

more than 350 km) for a period of six hours or more¹. Maddox did not place upper limits on his MCC definition, but his summary of 43 MCCs for 1978 ranged in diameter from 350 to 1000 km, with durations of 6-25 hours (except one >54 hours). Bosart and Sanders (1981) described the life history of a relatively small convective complex (at times, smaller than 300 km diameter), which originated over South Dakota, but caused disastrous flash floods over Pennsylvania four days later (well into the synoptic scale lifetime). The smallest entity noted in Figure 1.1 and the whole storm of Figure 1.2 fall short of the minimum dimensions given by Maddox, as do two particularly severe storms to be discussed in Section 1.3 (see Figures 1.7c, 1.8c). These smaller severe storms are more the 'normal' for Alberta and other High Plains² regions, and yet many of them develop their own mesoscale circulation (Lemon and Doswell, 1979).

Since there is apparently no clear relation between complex size, intensity, and duration, it makes sense to have at least two classifications of MCC. The Maddox MCC will therefore be called a Sub-synoptic Scale Convective Complex (SSCC), if horizontal dimensions exceed 200 km. Mesoscale Convective Complexes (MCC) will be defined to range from small clusters of cumulus clouds of 20 km across, up to small thunderstorm complexes of 200 km. While satellite imagery such as Figure 1.1 provide no proof of the scale interaction concept, the fact that groups of entities at one scale tend to be organized into entities

¹ These dimensions and durations refer to the upper cloud shield with (satellite) infrared temperature $< -32^{\circ}\text{C}$.

² The High Plains are loosely defined as the broad region in the lee of the Rocky Mountains, generally above 500 m elevation and extending from Texas northwest to Alberta.

at the next larger scale, suggests a strong relation between adjacent scales. Diagnostic studies, such as the four-day storm complex documented by Bosart and Sanders (1981), provide direct evidence of the interactions between different scales of motion.

1.3 SEVERE STORM RESEARCH: TRENDS AND CONTROVERSIES

Severe storm research embarked on two main courses during the 1950s and 1960s. One of these focussed its attention primarily at the cloud scale, the physics of cloud and precipitation particles, and how these might be altered to change the type and amount of precipitation at the ground. Out of this was born the science of weather modification. The other course of research was more interested in how synoptic scale processes modify the severe storm environment, and how the effects might be predicted. Because the two approaches focussed on almost opposite ends of the scale spectrum, they pursued virtually independent courses, but also developed apparently conflicting ideas as to cause and effect of convective storms.

In reality, recent mesoscale analyses are revealing that most of this conflict is unnecessary, and that, for the most part, the approaches are complementary. This section is presented in order to review some of the trends and controversies of the two approaches, since these are the sources of motivation for the current research, and why the author chose the several areas of focus which follow.

1.3.1 Controversy: Control "over" Versus Control "by"

After many years of severe storm research, most meteorologists agree that synoptic scale processes organize the mesoscale environment so that convective storms may form (Fawbush et al, 1951; Beebe and Bates, 1955; Ogura and Chen, 1977; Doswell, 1980). Some argue that the larger thunderstorms, once formed, develop some sort of autonomous mesoscale circulation, moving quite independently of, yet feeding back energy to, the mean synoptic flow. But others, especially those in forecasting roles, are of the opinion that the larger scale not only helps to initiate the severe storm, but continues to exert control throughout the storm's life cycle. Some specific case studies where the authors have demonstrated synoptic control over mesoscale systems include Ninomiya and Akiyama (1971) on east Asian rainfalls, Maddox et al (1978) on the Big Thompson flash flood, Bosart and Sanders (1981) on the Johnstown flash flood, and Bosart (1984) on a Texas coastal rainstorm. If this control is to be believed, energy exchanges between the scales involved are a logical consequence. The determination of the importance of the interactions between these scales of motion is one of the main objectives of STORM-Central, a mesoscale severe storm research project planned for 1989 over the central U.S. (Zipser, 1984). The estimated field budget of \$134 million attests to the importance accorded to scale interaction effects.

1.3.2 Priorities of Past Mesoscale Projects

Until recently, little research was directed towards downscale interactions between the synoptic and mesoscales, although, in view of what we have said above, the verification and detailed examination of the supposed energy exchanges is of the utmost importance. For example, in the first major severe storm field research effort, the Thunderstorm Project conducted over Florida and Ohio in 1947, most efforts were concentrated on the circulation, structure, and precipitation of the 'visible' storm and its immediate environment (Byers and Braham, 1949). The project initiated important studies in cloud physics and thunderstorm kinematics, including the description of the three stages of thunderstorm development. While these stages remain a valid simplified description of the cellular life cycle of storms, the real storm is now known to be far more complex and variable.

More recent research projects, while recognizing a role played by synoptic processes in severe storm development, still did not consider these important enough to include them in their experimental designs, other than for operational forecasting. Many of the High Plains experiments, through economic or other considerations, concentrated their efforts almost entirely on the storm itself. These include the National Hail Research Experiment of Colorado (NHRE, 1972-74), the Alberta Hail Project (AHP, 1974-present, and its predecessor, ALHAS, 1956-73), and even as recently as 1981, the Cooperative Convective Precipitation Experiment (CCOPE) of Montana. CCOPE did operate a small mesoscale rawinsonde network, but with only 5-7 sites spaced 70 km apart, analyses of interactions between synoptic and mesoscale processes using CCOPE data are not really feasible.

1.3.3 Research Trends: The "Capping Lid" and Synoptic Kinematics

The motivation for research into synoptic scale influences on severe storms has come not from the research community, but from severe weather forecasters, because of the necessity for improved forecasts of destructive and life-threatening storms. While the need for mesoscale predictions was recognized, progress in developing forecasting techniques was impeded by the lack of resolution of routinely available (upper air and surface) thermodynamic and wind data, which were of synoptic to sub-synoptic scale at best. Partly because of this, early research in this area concentrated on synoptic kinematics, climatological studies, and statistical forecast models, rather than on a theoretical approach.

One of the earliest significant advances here was the kinematic model developed by Beebe and Bates (1955), then of the Kansas City Severe Storms Center. They described how synoptic scale ascent might change a previously stable sounding into the type observed in the vicinity of severe thunderstorms (Figure 1.3). This conceptual model partially explained the role of the "capping inversion"³, or "lid", a thermodynamic signature 0.5-1.5 km above ground, which often precedes severe thunderstorms over the High Plains. The basic role of the lid is to temporarily trap moisture within the boundary layer, and to prevent

³ The capping inversion is an important severe storm signature throughout the High Plains regions from Texas to Alberta. The creation and breakdown of the inversion, including the role of topography, is important to the overall understanding of severe storms. This will become a focal point further on in this study. Suffice it to say that it is one of the links between storms of Texas/Oklahoma (the major subject region of this thesis), and similar ones in Alberta. This fact will facilitate comparisons and transferability of results between these two regions.

convection and latent heat release. Adiabatic cooling provided by synoptic scale ascent, often enhanced locally by orographic lift, eventually removes (or partially removes) the inversion, allowing a sudden release of the trapped energy, and often explosive growth of a severe thunderstorm.

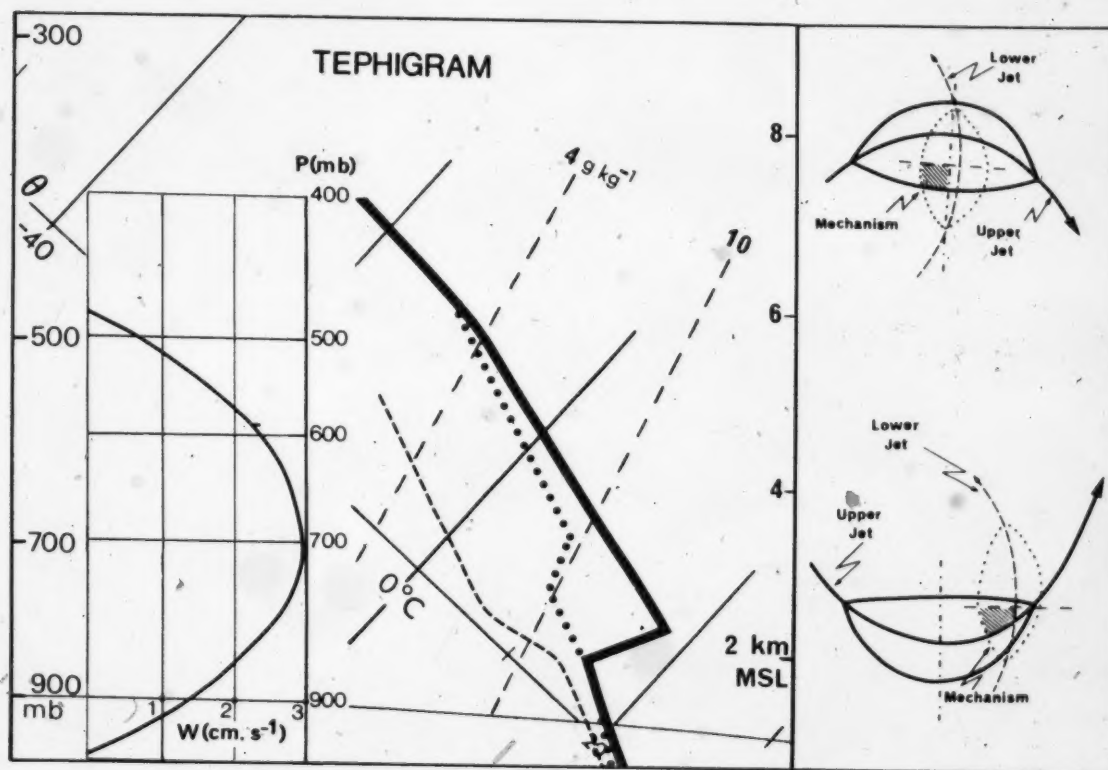


FIGURE 1.3: Average pre-tornado sounding showing temperature with capping lid (solid line), dew point (broken line), and mean tornado-vicinity temperature sounding (dotted line) following typical synoptic scale adiabatic ascent as shown. Typical upper/lower jet couplings for severe storm cases are also indicated. (After Beebe and Bates, 1955.)

Affirmation of the partial or complete removal of the capping lid in advance of storms was presented by Byers and Braham (1949), Beebe and Bates (1955), Darkow (1969), and others. This is further supported by observations of low-level convergence often preceding radar echoes of convection (e.g., Doneaud et al, 1983). The important factor here is that the ascent and cooling, whether it is synoptic scale ascent or otherwise, commences before the storm forms.

Closely related to this adiabatic cooling was Beebe and Bates' model of upper and lower jet structures (Figure 1.3) which assist in the release of convective instability. The left exit of a cyclonically curved upper jet, for example, is a preferred region for high level divergence with ascent at lower levels. The left entrance region will favour convergence, and therefore is likely to be associated with subsidence below this level. Upper level convergence would also favour cirrus clouds at these levels. Often, a low-level jet is also part of the overall structure, slightly downstream from the upper jet core and oriented across it. It will have a similar structure of divergence and convergence associated with it, with low-level convergence favouring ascent aloft.

These concepts are based on the equation for relative vorticity, expressed as a function of curvature and shear of the wind field (Beebe and Bates, 1955). Riehl et al (1952) described similar patterns for developing baroclinic waves based on the local change of vorticity. The intersection region where both jet structures favour ascent is the preferred region for thunderstorm formation (Beebe and Bates, 1955). In the presence of an anticyclonic curved jet, the favoured conditions for convection are slightly different. Then, divergence is most probable in

the right entrance region aloft, with the lower jet under or upstream from the upper jet core.

Uccellini and Johnson (1979) quantified these relations between wind jets and convective storms through diagnostic and numerical model results, and provided clues as to how the upper and lower jets are physically coupled through mass-momentum adjustments and transverse circulations within the exit region of the upper jet. Thus, the Beebe and Bates models still have relevance to the understanding and forecasting of thunderstorms some thirty years after their formulation.

The author has noted visual evidence of these effects from forecasting experience in Alberta. For example, a region where dynamic conditions favour severe storm formation, often has a pre-storm layer of altocumulus (AC) cloud, implying mid-level ascent. Convective storms tend to form near the upstream (rear) edge of an AC cloud layer. On satellite imagery, this is often referred to as the dry slot or dry intrusion of comma-shape cloud systems (e.g., Weber and Wilderotter, 1981; Millard and Carr, 1982). Late morning observations of altocumulus-castellanus (ACC) are an indication of mesoscale regions of even stronger ascent and existing mid-level instability. On the other hand, the appearance of a high layer of cirrostratus (CS) cloud (other than thunderstorm blow-off) with little other cloud present, is visual evidence of significant upper level convergence and probably mid-level subsidence, unfavourable conditions for convective development. CS cloud layers have the added effect of suppressing insolation, perhaps more so than AC layers which contain breaks between cloud entities. Another cloud type associated with weak or no convection over Alberta, is altocumulus standing lenticularis (ACSL). ACSL clouds over the

foothills frequently accompany a strong mid-level jet and near-laminar flow conditions, with little vertical motion at these levels except for cloud scale mountain waves which coincide with the lenticular clouds. The very strong shear associated with this structure also tends to suppress the updrafts vital to organized mesoscale convection.

1.3.4 Research Trends: Cloud Scale Processes

Despite initial directions provided by Beebe and Bates (1955) and others, the 1960s produced little new science that related thunderstorm formation to the synoptic scale. One major reason for the neglect of mesoscale research was the lack of surface and upper air data of sufficient resolution. To compound this, severe storm theoreticians, as well as those in field research, were focussing their efforts on the visible clouds and internal precipitation processes, rarely venturing outside the immediate storm environment. This concentration of effort resulted largely from the arrival of two important new technological tools during the late 1940s, weather radar, and aircraft equipped with sophisticated cloud physics instrumentation. The momentum of this research continued into the 60s and 70s, as faster, more compact computers which could even be taken into the field, allowed for more accurate processing of these data, and in greater quantities. The newer Doppler radars of the 80s guarantee a continued high level of research on the smaller scale in-cloud processes.

1.3.5 Research Trends: Synoptic Scale Climatology

It was therefore left to operational meteorologists such as Beebe and Bates to investigate mesoscale processes. Most of these efforts were necessarily directed towards developing prediction techniques, rather than formulating scientific theory. But even these modest efforts were hampered by the lack of mesoscale data resolution, depending almost exclusively on synoptic scale upper air and sub-synoptic scale surface data.

Synoptic climatology has been one means of investigating severe storms. In Canada for instance, Lowe and McKay (1962) produced climatological charts which imply that tornadoes over Manitoba and Saskatchewan occur primarily along a cold frontal trough, preceding the upper (500 mb) shortwave (Figure 1.4). Slightly contrary to this, a similar investigation of severe Alberta hailstorms (Longley and Thompson, 1965) provided mean charts which suggest that Alberta storms occur predominantly "behind" the low-level trough, in the cold baroclinic zone, but still preceding the upper trough (Figure 1.5). The author has confirmed (Strong, 1981) that both preferences are not incompatible (e.g., Figure 1.6), that major Alberta storms prefer "post-cold-frontal" situations, while Saskatchewan can have "cold-frontal" storms on the same day, associated with the same synoptic system. Thus, it is not too surprising that the average flow patterns of Figures 1.4 and 1.5 are quite similar. Because of this, one should avoid classifying convective events, particularly thunderstorms, simply as either frontal or airmass type. The formation or modification of airmasses by some underlying surface of water or land takes place under the influence of surface high pressure systems, the latter preceding and

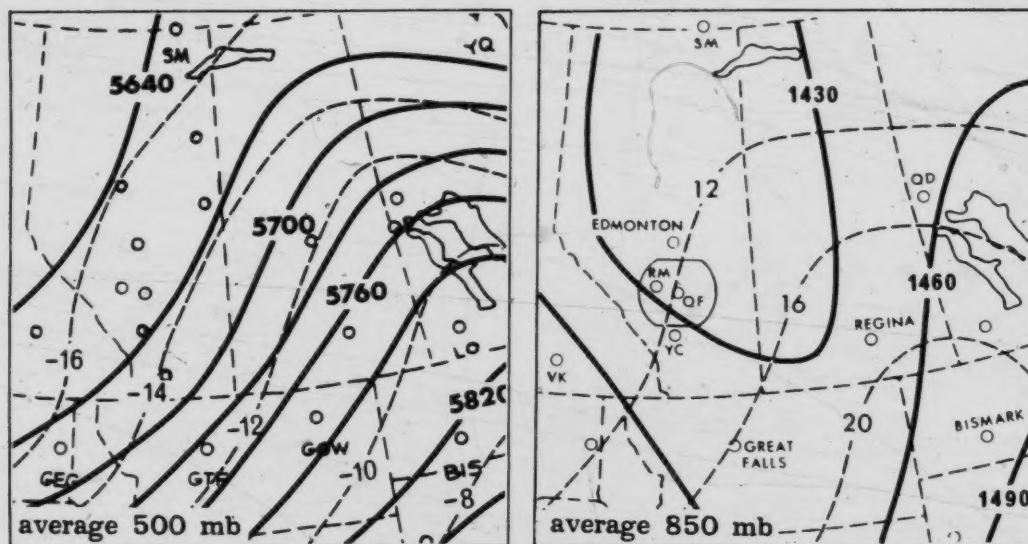


FIGURE 1.4: Average 500 and 850 mb flow patterns for Saskatchewan tornado occurrences (after Lowe and McKay, 1962).

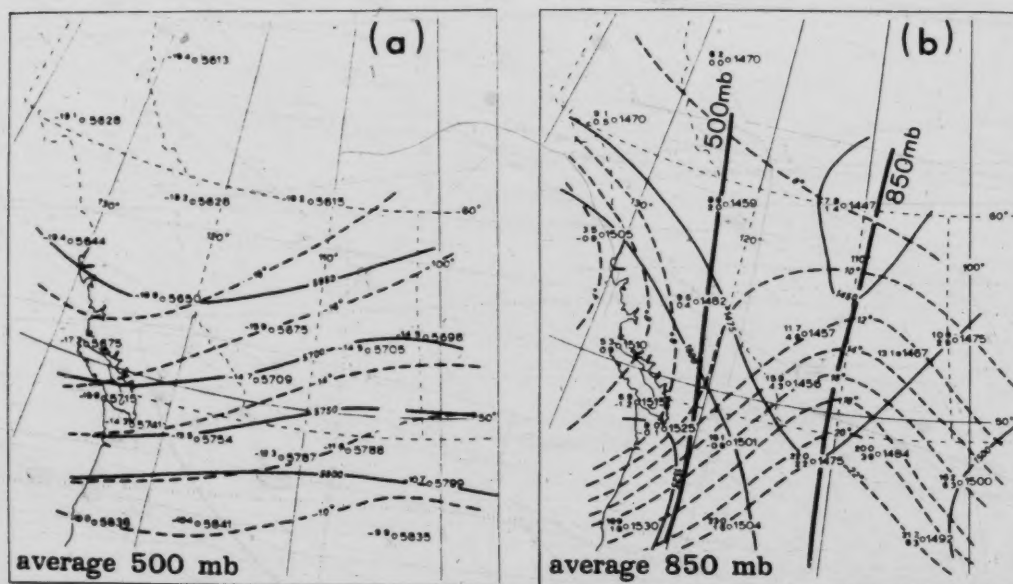


FIGURE 1.5: Average 500 and 850 mb flow patterns for severe Alberta hailstorm occurrences (after Longley and Thompson, 1965).

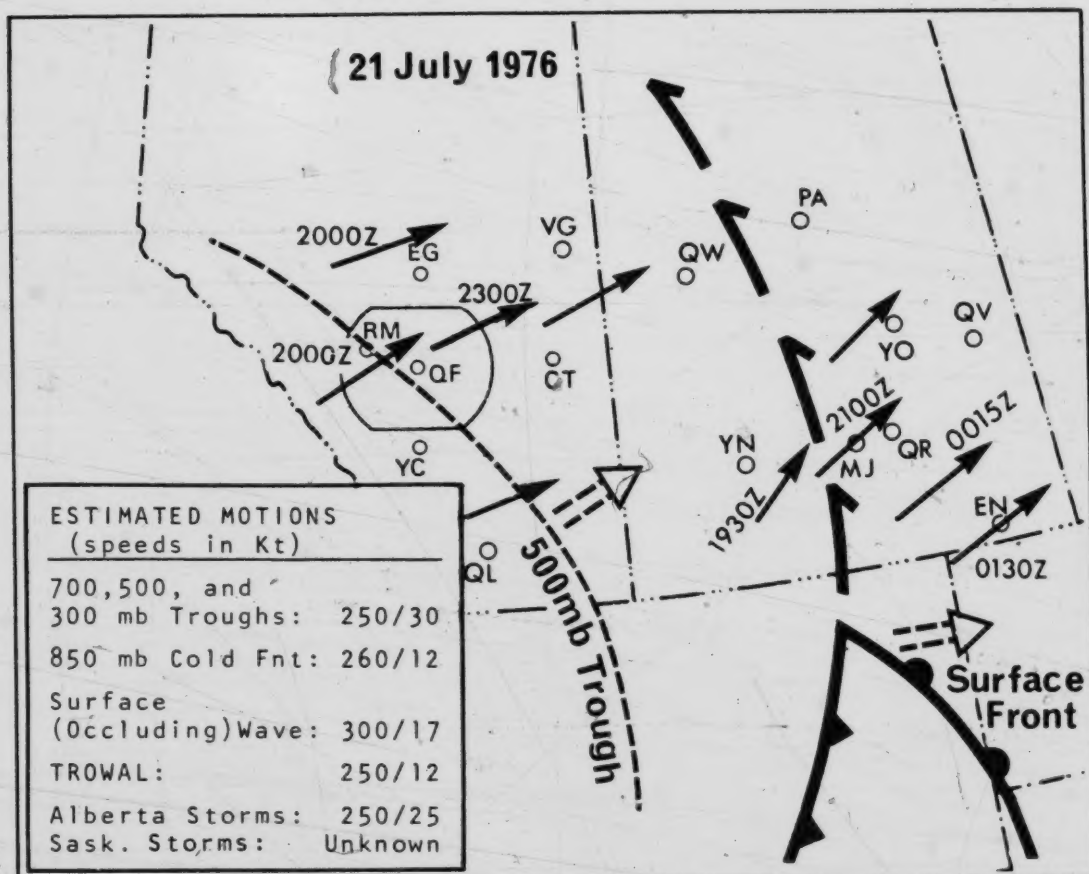


FIGURE 1.6: Schematic summary of known Alberta and Saskatchewan hailswaths in relation to the interpolated surface cold front and 500 mb trough at 2100 UTC, 21 July, 1976. Arrows indicate hailswath orientations while times are approximate known median hail times for each swath. (After Strong, 1981.)

being controlled by upper shortwave ridges. But severe storms rarely, if ever, occur over Alberta in these airmass situations, and neither can they be called strictly frontal (e.g., see Appendix Figure A1.4). A better approach to classification then, is to stratify convection according to the type of synoptic system, with fronts as qualifiers, if necessary (Strong, 1981).

1.3.6 Controversy: Low-level "Warm" or "Cold" Advection?

In either the Alberta or Saskatchewan storm cases, the important observation is that storms tend to form in locations where the atmosphere has already started to "cool at low to middle levels". This would seem to contradict the long-held view that thunderstorm formation regions are favoured by cold advection aloft, in conjunction with warm advection near the surface (Means, 1952; Newton, 1967; Ninomiya, 1971; Maddox, 1983). To be sure, warming at low levels would be the precursor, but if storm formation coincided most often with low-level warm advection, then one should also expect to see a greater frequency of warm-frontal storms rather than the observed cold-frontal activity.

The argument can be discussed from thermodynamic considerations; e.g., by examination of the Stability Tendency Equation (see Appendix III, Equation A3.1). This equation clearly shows that stability decreases with cold advection at all levels, but particularly when the advection increases with height. Some confusion arises from not distinguishing between warming caused by dynamic effects (both horizontal advection and subsidence), and that of diabatic heating. The former causes an increase in potential instability, except in the boundary layer, where these dynamic effects are suppressed by the lower boundary and by friction, creating a stable layer which also tends to suppress initiation of convection. On the other hand, diabatic heating tends to offset the latter, and is therefore an indisputable factor in storm formation which must be isolated from the two other local heating factors. Alberta statistics (Appendix III) clearly show that low- or mid-level cooling usually occurs prior to storm formation, and continues through its life cycle. Warm advection may well occur during this

period, but as is shown in Chapters 4 and 5, it is more than compensated for by adiabatic cooling due to ascent.

1.3.7 Research Trends: Forecast Techniques

The urgent need for forecast products sometimes obscured useful scientific results. For example, Longley and Thompson (1965) stated that "among the variables examined, no necessary and sufficient conditions were isolated by which one could distinguish a day with severe hail from days with either minor or no hail". They saw little significance in the patterns of Figures 1.4 and 1.5. Still, forecast techniques such as the composite analysis methods of Miller (1972), evolved from similar types of synoptic studies, and with moderate success. During the 1970s, more objective statistical applications of these concepts were developed. One fairly successful example is the Synoptic Index of Convection (Sc4), developed by Strong (1979).

The derivation and procedural use of the Sc4 is outlined in Appendix III. Briefly, it is computed from synoptic scale upper air variables and sub-synoptic scale surface data. The Sc4 gives reliable predictions of storm formation location, and its maximum convective intensity and relative size at maturity (Strong and Wilson, 1981, 1983ab). An objective procedure which combines satellite, radar, and surface moisture analyses with the Sc4, then provides considerable skill to forecast mesoscale storm features out to 18 hours (Strong and Wilson, 1983c). Often, the Sc4 correctly predicts storms where other techniques fail, and it gives correct storm motion (by prediction of new cell formations) in cases where it might appear that the storm had developed

its own circulation independent of the large scale flow. Two examples (from Strong and Wilson, 1983b,c) which illustrate the advantages of the Sc4 are summarized next. It will be shown how the examples have important implications for scale interactions.

1.3.8 Applications of the Synoptic Index

The Calgary, Alberta storm of 28 July 1981 (Figures 1.7e,f) was an intense sub-synoptic scale convective complex, which initially formed over the southwest quadrant of the Alberta Hail Project operations area (AHP), indicated by the quasi-circular region over southern Alberta. Its severity was marked by two deaths from lightning and damage claims well in excess of \$100 million. Its formation behind the surface frontal system but preceding the associated upper trough (Figures 1.7a,b) was typical of post-cold-frontal Alberta storms. Its subsequent motion to the southeast, fully 90 degrees to the right of the mid-level flow direction, was atypical, since the majority of storms follow the mean upper wind direction to within 20 degrees or less (Newton and Fankhauser, 1975).

The Sc4 (Figure 1.7c) and forecast procedure outlined in Appendix III indicates storm initiation along the foothills region of southwestern Alberta, where there was negative advection of Sc4. Maximum convective intensity should have occurred over or near the southwest quadrant of AHP. A region of maximum surface humidity extending southeastward from this location (not shown), suggests that new cells would have formed rapidly on the southeast side of the storm complex. Hence, there should have been rapid storm propagation towards

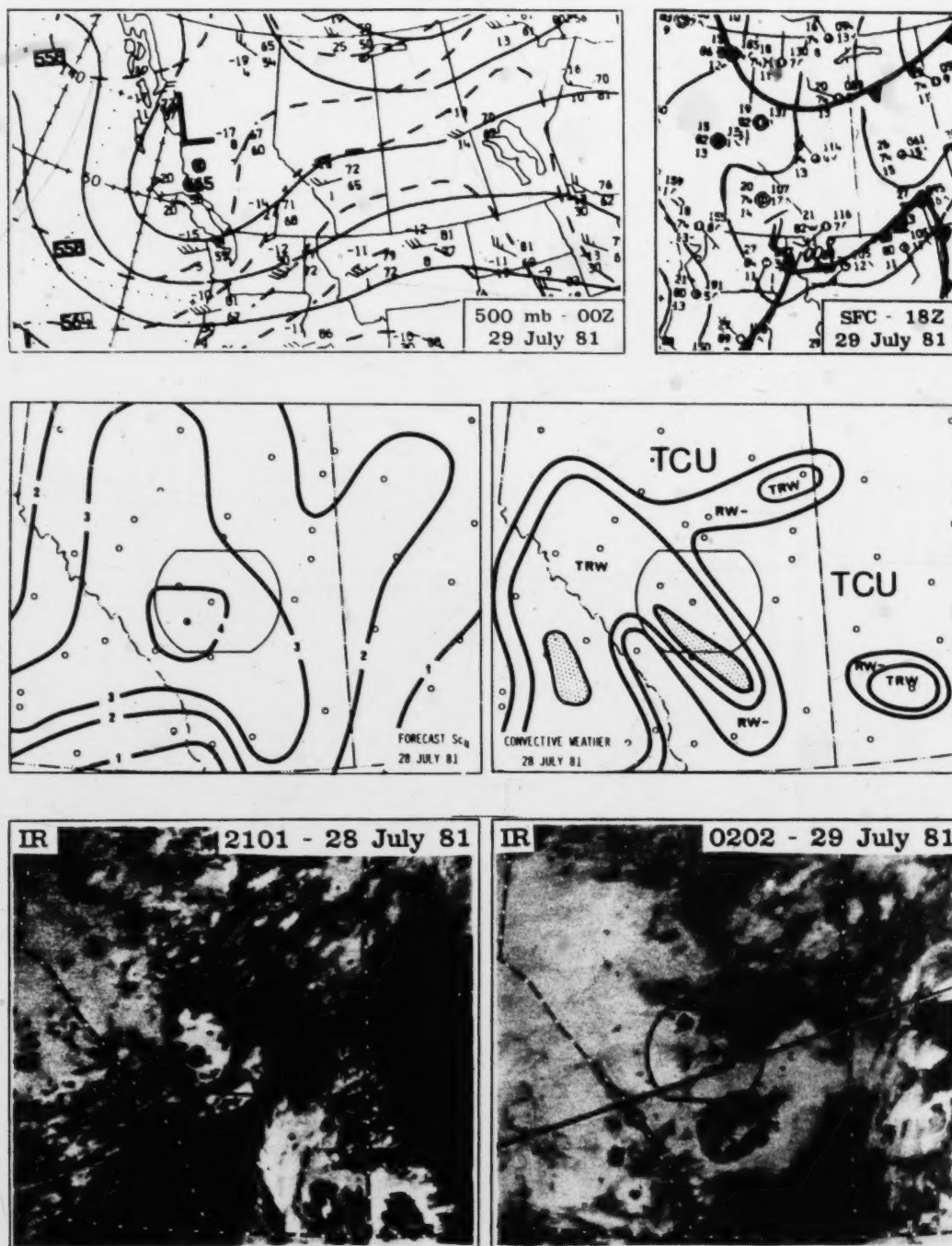


FIGURE 1.7: The Calgary \$100+ million hailstorm of 28 July, 1981: (a) 500 mb and (b) surface analyses at 2400 UTC; (c) 1500 UTC forecast Sc4 map; (d) diagnostic convective weather composite for 1700-1200 UTC, from surface weather observations, AHP hail reports, and satellite imagery; NOAA-6 enhanced infrared satellite image for (e) 2101 UTC, (f) 0202 UTC, 28-29 July. (After Strong and Wilson, 1983b.)

the southeast, as indeed did occur. Most of the other lobes extending out from the Sc4 maximum (Figure 1.7c) were also verified in the convective weather composite summary (Figure 1.7d).

The Fort Collins, Colorado storm of 30 July 1979 (Figure 1.8); is presented for several reasons: (1) it illustrates that the Synoptic Index of Convection, based on statistics for Alberta, can be applied elsewhere with reasonable confidence. (Later, it will be applied to the main case study analysis of this thesis.) (2) This storm also exhibited an abnormal motion relative to the mean upper wind directions. (3) Fritsch and Rodgers (1981) state that this storm, and particularly its motion, could not have been reliably forecast. (4) The storm falls outside the Maddox (1980) definition, yet its severity (12-cm hail in Fort Collins, one death, and \$20 million damage) warrants a special classification; in this case, a Mesoscale Convective Complex according to Table 1.1.

The synoptic situation (Figures 1.8a,b) bears some similarity to the Alberta case, in that the main surface front had passed Colorado, while a weak upper shortwave was still upstream (northwest Wyoming). A north-south line of mesoscale complexes had formed along the eastern foothills of the Rockies by 2124 UTC (Figure 1.8e), including one over southeast Wyoming, labelled 'S1' on the satellite image⁴ and

⁴ The convective nature and general intensities (S-severe, M-moderate, W-weak) labelled on the satellite image were inferred from hourly surface reports and information presented by Fritsch and Rodgers (1981) including radar composites.

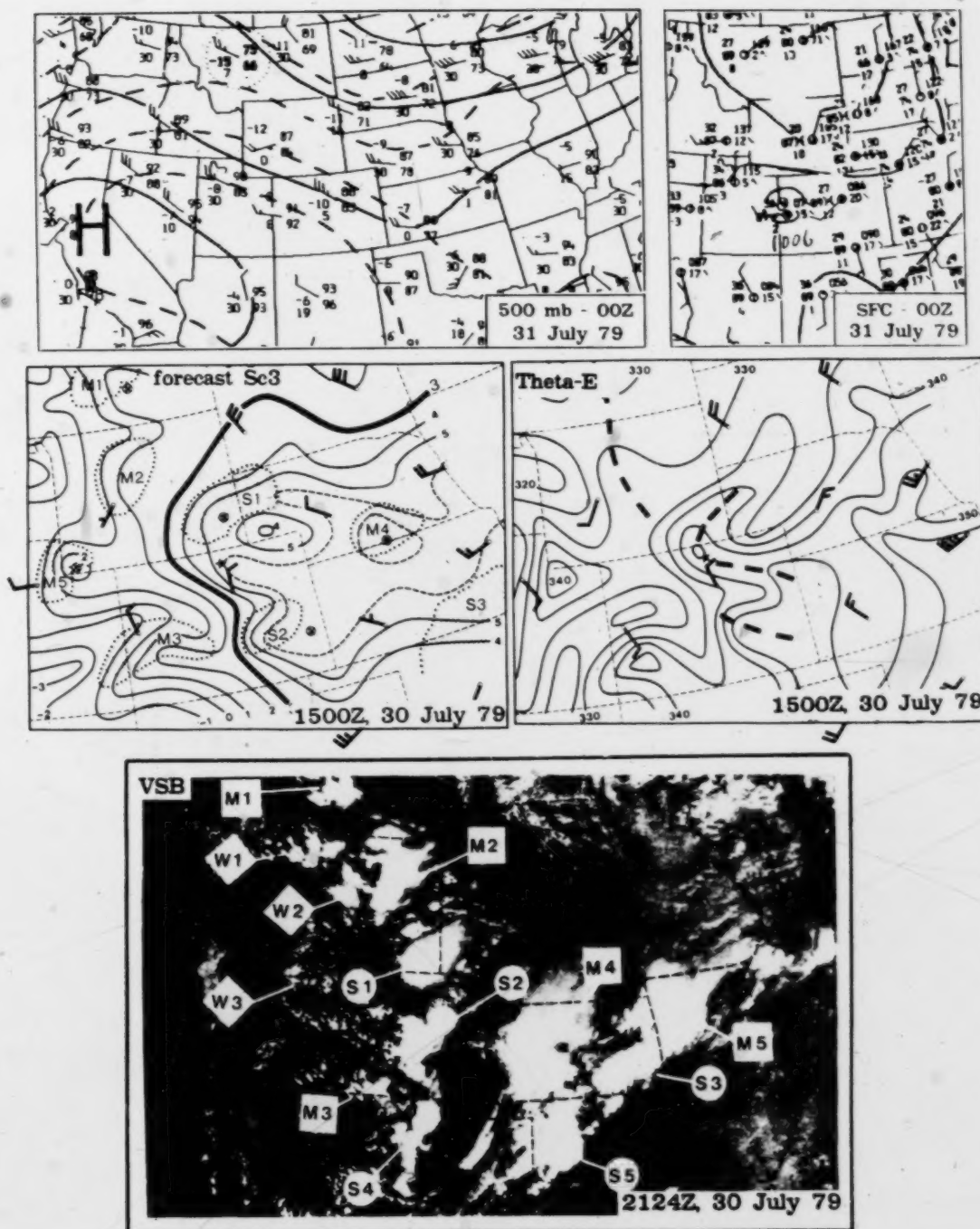


FIGURE 1.8: The Fort Collins, Colorado hailstorm of 30 July, 1979: (a) 500 mb and (b) surface analyses at 2400 UTC; (c) 1500 UTC forecast Sc3, 1200 UTC 700 mb winds, and predicted convective complexes (according to procedures outlined in Appendix III); (d) 1500 UTC surface Theta-E analysis and 1200 UTC 850 mb winds; (e) NOAA-6 visible satellite image for 2124 UTC, with convective complex intensities indicated as S-severe, M-moderate, and W-weak. (After Strong and Wilson, 1983b.)

corresponding Sc3 forecast⁵. These storms then moved eastward with the general upper air flow, with the exception of the Wyoming cell. This storm was moving straight south from 1900 until 2300 UTC when it struck Fort Collins. Thereafter it veered eastward with the other complexes.

Using the Sc4/Sc3 operational procedure outlined in Appendix III, one would forecast convective complex initiation over the regions of strongest negative Sc3 advection. Those regions within a +3 contour (the "severe storm threat region") would have a high probability of being severe (S1, S2, and S3 in Figures 1.8c and 1.8e), while outside the threat region, one would anticipate such regions to have weak to moderate storms (M1, M2, M3, and M5 in Figure 1.8c), depending on the values of Sc3. On the forecast Sc3 map, one should also predict the complex M4, since this region contained the highest contour value, +7. However, it was down-graded to moderate simply because no significant Sc3 advection is apparent there. The Sc3 forecast and satellite verification speak for themselves. The Sc3 over-forecast the complex labelled M5 (Figure 1.8c) over the mountains east of Salt Lake, where only weak cumulus cells occurred (W5 in Figure 1.8e), while M4 was predicted somewhat too far north. All other moderate to severe complexes within the domain of the data used were predicted rather well.

For storm motion, experience with the Sc4 (or Sc3) has shown that severe storms, once formed, tend to move towards the line of maximum Sc4. In this respect, the Fort Collins (S1) storm might have moved

⁵ One of the Sc4 input variables (the EGDEX) is normally not available for applications outside the Canadian Prairies, as it is dependent on a climatological diurnal adjustment to surface wet-bulb potential temperature. In such cases, the index is computed with only three variables (an Sc3), with a slight decrease in reliability.

either east or south, since higher values of Sc_3 lie in both directions, while the region directly southeast contains a minimum. However, for short-term (1-6 hour) forecasts once a storm is in progress, new convective cells will form where thermodynamic instability is greatest. A good measure of this instability is surface equivalent potential temperature (Θ_e , Figure 1.8d). In this case, 1500 UTC Θ_e values (Figure 1.8e) are $<335^\circ K$ to the east, but are some $15^\circ K$ higher to the south, with a maximum value ($350^\circ K$) near Fort Collins. Based on this, new cells would be expected to form rapidly on the south side of the main storm, producing the apparent southward propagation of the whole complex. Θ_e analyses have proven valuable for such short term forecasts in Alberta.

1.3.9 Controversy: Predictability and Independence of Severe Storms

Fritsch and Rodgers (1981) stated that "as convective systems grow larger, their impact on their environment significantly increases, and in some instances these systems even appear to detach themselves from any discernible travelling wave or synoptic forcing that may have initiated and supported them". This is an opinion which has been presented by others, and while such events may be possible, it is obviously not so in this instance, since the forecast is based on 1500 UTC data, several hours before the storms formed. In this case, the storm formed due to predictable synoptic scale processes, and the sub-synoptic scale moisture field measured during the pre-storm period. For a short period, the storm's motion was controlled mainly by this low-level moisture field, which itself was organized predominantly by

the synoptic scale, with mesoscale concentrations of moisture due to convergence within river valleys or other sub-mesoscale topographic features. In this case at least, the Sc3 forecast and Theta-E analysis would have proven Fritsch and Rodgers incorrect on the non-predictability of the storm formation, intensity, and motion. One must wonder if such is the case for other severe storms where the analyst has detected no other logical predictive capability. It raises questions concerning how well we make use of existing data.

Clusters of convective storms tend to deflect to the right of the mean environmental winds because of the systematic pattern of new cell formation on the upwind side (Newton and Fankhauser, 1975). The degree of this deflection seems to be in direct proportion to the storm's intensity and its rate of development, both being functions of the degree of conditional instability present and the balance of supply and demand for water vapour. These factors are preset by synoptic processes acting on the mesoscale in conjunction with local topography. Thus, net storm motion is such as to intercept the required supply of water vapour. The above applications of the Sc4 to the extreme, right-deflecting Calgary and Fort Collins storms are important because they illustrate that these mesoscale processes are deterministic, that the formation and motion of severe convective complexes can be predicted, even, to a large degree, from existing synoptic and sub-synoptic scale data. This demonstrates as well that such storms do not behave independently of the larger scale processes.

1.3.10 Technological Influences on Severe Storm Research

Severe storm research had taken major steps forward by the 1970s. While there was a concentration of research on the visible, mature severe storm, two other developments started to direct some focus back to the storm environment, and more recently to the subject of scale interactions. These were the advent of high-resolution satellite imagery, and the operation of several mesoscale upper air networks.

Satellite imagery have revealed a degree of organization to severe convective storms, previously attributed only to synoptic scale features. This includes the recognition of large mesoscale convective complexes (Maddox, 1980). When a number of complexes occur together within a large region, they also show a tendency for synoptic scale organization and influence (Figure 1.1).

The mesoscale networks, notably those of the National Severe Storms Laboratory (NSSL, Oklahoma, 1966-67), and the Severe Environmental Storms and Mesoscale Experiment (SESAME, Oklahoma/Texas, 1979), were designed to study meteorological aspects of the severe storm and its immediate environment, rather than interactions with the synoptic scale specifically. However, they have also provided a useful data base for scale interaction studies.

NSSL data revealed mesoscale organization in the pre-storm environment, which was attributed primarily to convectively-induced surface convergence. Realization of the tremendous amounts of latent heat produced by such storms provoked speculation as to its feedback effect on the larger scale environment. These considerations prompted the design of the SESAME network, which, for the first time, involved

simultaneous three-hour releases from the surrounding synoptic as well as the mesoscale network.

1.3.11 Research Trends: Mesoscale Modeling

Mesoscale models, initialized with pre-storm environmental data, have three important potential uses. First, they provide one method of testing scientific hypotheses concerning interactions with the mesoscale environment. Second, if one can successfully simulate convective environmental conditions using mesoscale data for initialization and for ground-truthing, then one can vary the model input to simulate other conditions for which one lacks the necessary mesoscale data. Assuming that the model performs well for these applications, the third use, as a real-time forecasting tool, is rather obvious. Pielke (1984) provides an excellent review of current mesoscale models, many of which are fast approaching this potential.

1.4 OBJECTIVES OF THIS STUDY

While the NSSL and SESAME data have helped to widen the scope of mesoscale research, the trends of previous field activities remain, with a continued emphasis on the mature, severe storm. Even those mesoscale studies which have investigated scale interactions, have done so by concentrating on the feedback of energy from the severe storm to the larger scale. The concept of downscale effects, particularly on the pre-storm environment, the forecaster's main concern, has been neglected in most studies.

Forecasting techniques such as the Synoptic Index, rely on synoptic and sub-synoptic scale measurements, which are unaffected by later convective events. There remain many in severe storm research who doubt any current predictive skill for mesoscale convective events. However, the success shown by the Synoptic Index in predicting such events, may well be a demonstration of the importance of downscale effects on the pre-storm environment. Further, it raises the question as to whether we are making the best possible use of existing data sources.

The foregoing trends and controversies, and the author's statistical results and forecasting experience, have determined the major scientific objectives of this research, which are:

- 1) to formulate a hypothesis which links scale interactions to the formation of convective storms, especially over Alberta;
- 2) to test this scale interaction hypothesis with field evidence;
- 3) to better utilize existing data sources for forecasting such storms;
- 4) based on the hypothesis test, to search out new techniques of recognizing and forecasting these storms;
- 5) to provide a basis for a "total" conceptual model of thunderstorms, as it applies to Alberta and elsewhere, one which looks beyond the visible convective processes and includes the concept of scale interactions;
- 6) and finally, to make recommendations for future field research which will lead to a better understanding of scale interactions, and ultimately to much improved forecasts of convective events, including quantitative convective precipitation forecasts (QCPFs).

1.5 SUMMARY

A theory of scale interactions in the synoptic to planetary scale range was developed during the late 1940s and early 50s by Charney, Starr, Rossby, Eady, Lorenz, and several others. The theory evolved from considerations of potential and kinetic energy budgets, first discussed by Margules (1903). In reviewing some of the qualitative evidence for scale interactions at the mesoscale, the controversial nature of the concept has also been emphasized, especially with regard to how synoptic processes might influence the pre-storm environment and even the mature storm (downscale effects), or be affected by the storm (feedback effects).

In the past, severe storm research has been generally directed towards understanding the visible entity, the thunderstorm itself. Because of this focus on the cloud scale and smaller scales, the pre-storm environment, without any "interesting" clouds, has been largely neglected. Consequently, the scientific community has been slow to recognize and diagnose any interactions between mesoscale storms and much larger atmospheric processes. This trend also influenced technological development for the field, with emphasis on radars and cloud physics instrumentation with which to measure the visible storm alone. The forecaster is forced to use larger scale data sets, but yet must make predictive inferences regarding the mesoscale. As a result, he has accepted a concept of interacting scales more easily and sooner (e.g., Fawbush and Miller, 1951), but without a deep scientific understanding of the processes involved.

In this respect, the operational community can be credited with the first steps towards scale interaction research on severe storms, even though most of this has been either of a qualitative or of a statistical nature. Among these steps, we include the kinematic and climatological studies of Beebe and Bates (1955), Lowe and McKay (1962), Longley and Thompson (1965), Miller (1972), and more recently, Maddox (1980). The Synoptic Index of Convection, and the forecast technique outlined in Appendix III, represent a statistical application of the scale interaction concept, and have proven useful for predicting the initiation region, size, motion, and the location of maximum intensity of convective storms.

The research trend has changed during the last few years with the recognition of large, sub-synoptic and mesoscale convective complexes. These exhibit organization and circulation patterns not unlike synoptic processes in many instances. Detailed mesoscale analyses of the pre-storm and storm environment are now possible through the use of data sets such as those provided by the 1979 SESAME cases. The STORM-Central Experiment (Zipser, 1984) will provide the tools to resolve much of the controversy concerning scale interactions, and improve our forecasting capabilities at the mesoscale.

The fact that environmental winds flow through, as well as around thunderstorm systems, gives new meaning to Rossby's observation (quoted at the head of this chapter) regarding scale interactions and exchanges of energy between disturbances. Interactions between all scales are manifested in the atmosphere through local changes in the global energy balance. Nature maintains this energy balance through exchanges of potential and kinetic energy, including that of thunderstorm systems.

Before reporting the results of current efforts towards the goals of Section 1.4, it is therefore necessary to review some of the theoretical and diagnostic background on energetics. Chapter 2 is devoted to this end.

*"Big whirls have little whirls
that feed on their velocity,
And little whirls have lesser whirls
and so on to viscosity."*

L.F. Richardson (1922)

CHAPTER 2.

THE ENERGETICS OF SCALE INTERACTIONS

Up to this stage, one might accept the concept that interactions occur between synoptic processes and the convective storm environment. These interactions must also involve exchanges of energy between the scales. It has also been noted, here and in the literature, that there exist mesoscale circulation patterns associated with convective storms which are similar to those of the synoptic scale. Therefore, following the logic in Richardson's verse above, since atmospheric energy exchanges are dominated by those at the synoptic and planetary scales, it is important to examine these exchanges before investigating those at smaller scales. At the same time, the dynamics of large scale interactions are better understood than those at the mesoscale, mainly because of the availability of a much more extensive and global data base. With these points in mind, this study will first review the known features of large scale circulation systems and atmospheric energetics.

2.1 LARGE SCALE ENERGETICS THEORY

Long before scale interactions were recognized, it was known that synoptic disturbances, alternately referred to as baroclinic waves or extra-tropical cyclones, experience an increase in kinetic energy as they intensify. Three schools of thought emerged as to how this increase in kinetic energy occurred. The original theory, attributed to Margules (1903), was that the kinetic energy (KE) of a cyclone increased through an in situ conversion of available potential energy (APE)⁶ by an overturning or readjustment of airmasses of different density.

During the 1940s, revelations of the tremendous kinetic energy carried by the upper tropospheric jet streams led to renewed interest in energy conversion processes. Thereafter, the Margules' theory ran into trouble when Starr (1948), Rossby (1949), and others showed mathematically that single cyclones cannot be even approximated as closed systems. Rossby explained that this is so because it is impossible to have a KE source in a closed system without also having sinks of KE, independent of frictional sinks. Using a scaling argument, he showed that frictional dissipation alone does not account for the total energy sink observed in single cyclones. Rossby's earlier quote and simple observation (beginning of Chapter 1) makes the argument even clearer. Since the same upper winds blow through and around convective clouds and storms of various scales, one may also conclude that similar

⁶ Margules (1903) recognized a maximum attainable gain of kinetic energy through the adiabatic redistribution of total potential energy, and called it the "available kinetic energy" (AKE). Lorenz (1955) redefined this as the "available potential energy" (APE), being equal to the maximum amount of the total potential energy available for conversion into kinetic energy under adiabatic redistribution of mass.

energy exchanges take place in this instance. Scaling arguments show that the environmental energy flowing into the storm region is of the same order of magnitude as the thermodynamic energy generated by the storm through latent heat release.

Two alternate theories to explain the intensification of cyclones were developed around this time. Charney (1947) and Eady (1949) independently showed that APE of the zonal (planetary mean) flow was converted to (synoptic scale) KE in developing cyclones. But this appeared to be in conflict with Kuo (1949), who demonstrated how the KE of the cyclone could come directly from KE of the planetary mean flow. In one of the earliest diagnostic studies of energetics, Starr (1953) introduced a new complexity. By using hemispheric wind observations, he suggested that the synoptic scale feeds back at least some of its KE to the mean planetary flow, thereby maintaining the latter against frictional dissipation.

These problems were partially resolved by Lorenz (1955), who by partitioning the APE and KE budget equations into zonal and eddy components, showed that both the Charney/Eady and Kuo theories were possible. Lorenz also emphasized the distinction between frictional and eddy dissipation, by showing that friction alone could not account for the observed energy sink. Consequently, he inferred that there must exist intermediate eddies which hold the balance of energy between the observed synoptic scale and friction. Eliassen (1966) removed any remaining conflict by proving that there were both APE and AKE at both scales (planetary and synoptic) for conversion to perturbation energy of both types, and these could be redistributed either upscale or downscale. Thus, Margules' (1903) original theory was not entirely

incorrect either, since synoptic scale APE could presumably be converted to KE at the same scale. McInnis and Kung (1972) provided the first results showing that these conversions of energy are applicable down to the mesoscale range as well.

Following the lead of Lorenz (1955, 1960), diagnostic studies of energetics became quite popular, particularly after much of this theoretical controversy had been resolved. The purpose of the diagnostic studies has been to determine the relative importance of the various terms in the energy budget equations.

2.2 KINETIC ENERGY EQUATION FOR THE MEAN FLOW

The Eulerian form of the KE budget can be derived directly from the equation of motion for the horizontal wind,

$$\frac{d\bar{\mathbf{V}}}{dt} = -f\hat{\mathbf{k}} \times \bar{\mathbf{V}} - \bar{\nabla}\phi - \bar{\mathbf{F}} \quad (2.1)$$

where $\bar{\mathbf{V}} = u\hat{\mathbf{i}} + v\hat{\mathbf{j}}$, the horizontal wind,
 $f = 2\Omega \sin\varphi$, the Coriolis parameter,
 and $\phi = gz$, the geopotential.

$\bar{\nabla}$ is the horizontal gradient operator, and $\bar{\mathbf{F}}$ represents the net frictional force per unit mass. This includes molecular viscosity (a few cm above the earth's surface, which can be neglected), but is eddy viscosity otherwise.

To express (2.1) in an energy form, we take its scalar product with $\vec{V}$, and then expand the total derivative of K to give:

$$\begin{aligned}\frac{dK}{dt} &= \frac{\partial K}{\partial t} + u \frac{\partial K}{\partial x} + v \frac{\partial K}{\partial y} + \omega \frac{\partial K}{\partial p} \\ &= \frac{\partial K}{\partial t} + \vec{V} \cdot \nabla K + \omega \frac{\partial K}{\partial p}\end{aligned}$$

The second term in (2.1) disappears, since $\vec{V} \cdot (-f\hat{k} \times \vec{V}) = 0$; that is, the Coriolis component is zero, since it acts normal to the motion and does no work.

We now have

$$\frac{\partial K}{\partial t} = -\vec{V} \cdot \vec{\nabla} K - \omega \frac{\partial K}{\partial p} - \vec{V} \cdot \vec{\nabla} \phi - D \quad (2.2)$$

where $D = \vec{V} \cdot \vec{F}$ is the dissipation of energy through eddy viscosity,

$$K = \frac{1}{2}(u^2 + v^2) \quad (2.3)$$

Applying the Hydrostatic Equation, and the definition of geopotential,

$$\frac{\partial \phi}{\partial p} = -\alpha \quad (2.4)$$

along with the Continuity Equation in isobaric coordinates,

$$\frac{\partial \omega}{\partial p} = \vec{\nabla} \cdot \vec{V} \quad (2.5)$$

to the second and third terms of (2.2),

$$\begin{aligned}\vec{V} \cdot \vec{\nabla} K + \omega \frac{\partial K}{\partial p} &= \vec{V} \cdot \vec{\nabla} K + \frac{\partial(\omega K)}{\partial p} - K \frac{\partial \omega}{\partial p} \\ &= \vec{V} \cdot \vec{\nabla} K + \frac{\partial(\omega K)}{\partial p} + K(\vec{V} \cdot \vec{\nabla}) \\ &= \vec{V} \cdot \vec{\nabla} K + \frac{\partial(\omega K)}{\partial p}\end{aligned}$$

so that, on integrating over a volume of the atmosphere, (2.2) becomes

$$\int \frac{\partial K}{\partial t} = \int \underbrace{-\nabla \cdot \mathbf{V}K}_{(\text{HFK})} + \int \underbrace{-\frac{\partial(\omega K)}{\partial p}}_{(\text{VFK})} + \int \underbrace{-\mathbf{V} \cdot \nabla \phi}_{(\text{CK})} - \int \underbrace{D}_{(\text{D})} \quad (2.6)$$

where the integral $\int = \frac{1}{gA} \int_p \int_y \int_x dx dy dp$

Kung and Tsui (1975) express the mean value of a dummy variable X over an analysis area A on an isobaric surface as

$$[X] = \frac{1}{A} \int_A X dA \quad (2.7)$$

so that (2.6) can be rewritten with the $\int$ replaced by $[]$ to

$$\left[\frac{\partial K}{\partial t} \right] = \left[-\nabla \cdot \mathbf{V}K \right] + \left[-\frac{\partial(\omega K)}{\partial p} \right] + \left[-\mathbf{V} \cdot \nabla \phi \right] + [D] \quad (2.8)$$

The terms in (2.8) represent, in plain language:

$$\left\{ \begin{array}{l} \text{Local rate} \\ \text{of Change} \\ \text{of KE} \end{array} \right\} = \left\{ \begin{array}{l} \text{Horizontal} \\ \text{flux of} \\ \text{KE} \end{array} \right\} + \left\{ \begin{array}{l} \text{Vertical} \\ \text{flux of} \\ \text{KE} \end{array} \right\} + \left\{ \begin{array}{l} \text{Cross-contour} \\ \text{Conversion} \\ \text{of APE} \\ \text{to KE} \end{array} \right\} - \left\{ \begin{array}{l} \text{Dissipation} \\ \text{of KE} \\ \text{(friction,} \\ \text{eddies, etc.)} \end{array} \right\} \quad (2.8a)$$

2.3 AVAILABLE POTENTIAL ENERGY EQUATION FOR THE MEAN FLOW

The Eulerian APE budget equation for the mean flow (see Smith, 1969 for derivation) can be similarly stated as

$$\int \frac{\partial A}{\partial t} = \int N \dot{Q} + \int \omega \alpha + \int -\nabla_p c_p N T V + \int \frac{c_p T}{p^\kappa} \frac{dp^\kappa}{dt} \quad (2.9)$$

(G) (CA) (HFA) (DP)

or:

$$\left\{ \begin{array}{l} \text{Local rate} \\ \text{of Change} \\ \text{of APE} \end{array} \right\} = \left\{ \begin{array}{l} \text{Conversion} \\ \text{APE} \\ \text{by diabatic} \\ \text{heating} \end{array} \right\} + \left\{ \begin{array}{l} \text{Conversion} \\ \text{APE within} \\ \text{the region to} \\ \text{KE either within} \\ \text{or outside region} \end{array} \right\} + \left\{ \begin{array}{l} \text{Horizontal} \\ \text{flux} \\ \text{Divergence} \\ \text{of APE} \end{array} \right\} + \left\{ \begin{array}{l} \text{Integrated Change} \\ \text{of ref. pressure} \\ (=0 \text{ when integrated} \\ \text{over whole} \\ \text{atmosphere}) \end{array} \right\}$$

(2.9a)

Pearce (1974) showed that for a closed system, APE generation ($G > 0$) occurs where the pressure and temperature are higher than the system average, while $G < 0$ occurs where pressure and temperature are below the average. Thus, according to Pearce, there should be an increase in APE within the warm sector region of cyclones.

The generation term (G) contains contributions from latent heat release, solar and terrestrial radiation (heating/cooling), and sensible heating, with latent heat being the primary source (Smith, 1980). The effects of latent heat release on the cyclone life cycle have been investigated quite extensively. Danard (1966) and Bullock and Johnson (1971) concluded that latent heat alone cannot initiate cyclone development, but that it can be an important energy source after the cyclone circulation is established. The same may well be true with regard to the initiation and maintenance of convective complexes.

Cho et al. (1979) found that, for limited areas, the released energy of tropical cumulus clouds adds a vertical component of vorticity comparable to that of the large scale vorticity field. However, this contribution was not large enough to support the generation or maintenance of tropical easterly waves.

The KE and APE budget equations, (2.6) and (2.9); are related through the terms (CK) and (CA) by

$$CA + CK = \int -\nabla p \cdot \phi V + \int -\frac{\partial \omega \phi}{\partial p} \quad (2.10)$$

The two RHS terms represent work done by the pressure gradient on the horizontal and vertical boundaries of the system, and are referred to as the boundary work terms (Smith, 1970).

2.4 PROBLEMS OF ENERGY BUDGET COMPUTATIONS

Most diagnostic studies of energetics have employed the Eulerian forms of the equations, as given by (2.8) and (2.9). Case studies using Lagrangian formulations have been presented by Eddy (1965), Johnson (1970), Vincent and Chang (1975), and Chen and Bosart (1977). These studies, following single weather systems, assume that the system is closed, and thereby neglect any system-to-system energy interactions.

In practice, it is difficult to interpret APE budgets for limited regions of the atmosphere. The efficiency of the atmosphere for converting PE to APE has been estimated at only 0.5%, with only 0.05% actually being converted to kinetic energy (Lorenz, 1955). The

conversion of potential to kinetic energy is necessary in order to compensate for frictional losses which may occur either within or outside of an isolated analysis region. Because of this, APE budget analyses for closed regions of the atmosphere have limited usefulness. However, Johnson (1970) circumvented this problem with open cyclonic systems to some extent, by partitioning APE into baroclinic and barotropic components. The baroclinic partition represents the energy exchange if the system were closed, and is called the local APE. The most useful applications of APE budgets have been in general circulation studies of global or hemispheric dimensions. In these cases, the long-term mean conversion of APE to KE must balance the dissipation (Wiin-Nielsen, 1959).

One must be cautious when discussing individual values of terms in the budget equations, especially for APE budgets. For example, Smith (1980) reported rather large values of CA (-280 W m^{-2}), HFA ($+100 \text{ W m}^{-2}$), and DP ($+220 \text{ W m}^{-2}$) for cyclones, in comparison with G, the generation term (only 10 W m^{-2}). But he explained that this indicates that much of the APE is exchanged with the surrounding environment, so that the generation term has a more direct impact on cyclone evolution than the three larger terms.

Kung (1966, and personal communication) emphasizes that KE budgets can be computed directly from wind observations, the accuracy being dependent only on the resolution and accuracy of the wind and geopotential measurements within the open system. The various terms in (2.8) are usually computed as area averages. On the one hand, KE budgets have an advantage over PE budget computations, in that they can be applied with reasonable confidence to open portions of the

atmosphere, such as a single cyclone; i.e., the value for APE conversion to KE for KE budgets (CK in (2.8)) is dependent only on the terms within the analysis region. However, misinterpretations of KE budgets can occur because the dissipation term (D) is obtained as the (mathematical) residual in balancing the equation. The weakness here is that this term contains any observational errors as well as both frictional and eddy dissipation.

Kung and Tsui (1975) also point out that even without observational errors, (D) represents only the "net" sink or disappearance of KE "from the observed scale" through dissipation. Its value can even be negative, as in Kung and Baker's (1975) results for anticyclones, meaning a net input of energy from other (sub- or super-grid) scales to the observed scale of motion. Starr (1948) noted this possibility, commenting that (D) also embraces energy-producing systems. This emphasizes the fact that (D) is not a dissipation in the usual meaning of the word, but rather a disappearance/appearance of energy from/to the observed scale of motion. Of course, (D) also contains the errors of observation and computation. Using data from one or more large mesoscale networks nested within a uniform synoptic network, with all sites providing data with high temporal resolution simultaneously, it appears possible to at least partially resolve many of the problem terms by computing budgets at several scales of motion. The proposed STORM-Central networks (Zipser, 1984) would provide this opportunity.

Studies of energetics which have been averaged spatially and temporally, reveal certain fixed features of cyclone behaviour. But, as pointed out by Johnson (1970) and Smith (1980), hemispheric and time-averaged diagnoses of energy budgets often conceal more dramatic

energy transformations that occur in individual cyclones of limited horizontal dimensions. Individual deviations of the variables from mean values can range over two orders of magnitude. For example, Kung and Baker (1975) computed a mean seasonal value of dissipation for cyclonic disturbances of 6.3 W m^{-2} . On the other hand, comparative (cyclonic) case study values range from -1.6 W m^{-2} (Fuelberg and Scoggins, 1978) to $+59.1 \text{ W m}^{-2}$ (Chen and Bosart, 1977). Case studies of individual systems therefore illustrate the extremes of energy exchange which can and do occur in the atmosphere. In particular, it is imperative to know where and how, in relation to the storm, the conversions of APE to KE and KE dissipation are most significant. In doing this, one must recognize the problems of dealing with limited regions of the atmosphere, especially for APE budget studies. Some of the findings to date follow.

2.5 HIGHLIGHTS OF PLANETARY/SYNOPTIC SCALE DIAGNOSTIC STUDIES

Wiin-Nielsen (1959) published results from one of the earliest hemispheric diagnostic studies of energy conversions. During the winter and spring months, there was a positive conversion of APE to KE for each day of his two-month sample. Maximum conversions occurred for wave numbers 2 (planetary wavelength of approximately 20,000 km) and 6 or 7 (wavelengths of 5000-6000 km). He concluded that the longwaves received most of their KE from the shorter waves through non-linear interactions.

There followed a long series of diagnostic studies of planetary to synoptic scale energetics. These have been spearheaded by E.C. Kung and his associates at the University of Missouri, Smith and Vincent at Purdue University, and more recently, Fuelberg and his associates at

St. Louis University. Kung has studied annual (winter/summer) and diurnal (0000/1200 UTC) variations in the energy budget terms, variations with height, changes from continent to ocean, and meridional distributions. Some of the major observations of the large scale energetics studies are now summarized.

2.5.1 Distribution with Height of KE Budget Terms

Kung and Baker (1975) used five years of North American synoptic data and grouped the daily flow patterns into 11 different categories, including various cyclonic, anticyclonic, and intermediate flow types. Figure 2.1 shows the five-year average (Eulerian) vertical profiles for the terms of Equation (2.8), plus the vertical velocity in pressure coordinates for cyclonic and anticyclonic flow patterns. Some key results of this study can be summarized.

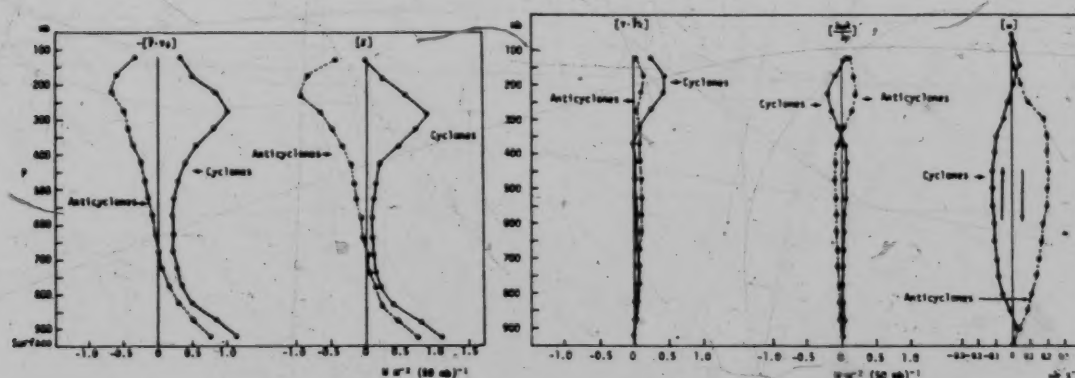


FIGURE 2.1: Mean synoptic scale cyclonic and anticyclonic vertical profiles of (a) cross isobaric generation and dissipation of KE, and (b) horizontal and vertical fluxes of KE and vertical velocity (after Kung and Baker, 1975).

- a) For cyclonic flows, peak values of mean cross-isobaric kinetic energy generation and dissipation occurred in the lower troposphere (below 800 mb) and at the jet stream level.
- b) For anticyclonic flows, the sign of the generation and dissipation terms tends to reverse in the upper troposphere. This indicates depletion of kinetic energy through (reverse) cross-isobaric flow. The negative value for jet stream dissipation is actually an energy source for anticyclones, representing a supply of energy from other (possibly the unresolvable sub-grid) scales of motion. The sign reversal appears to be necessary for the maintenance of the anticyclonic flow. The positive peaks of generation and dissipation in the boundary layer were similar to the cyclonic flow cases.
- c) The average results for all 11 categories show that the boundary layer generation and dissipation terms approximately balance each other, both for cyclonic and anticyclonic flows. There is a net generation of kinetic energy from cross-isobaric flow, and a net dissipation of kinetic energy to smaller scales. The advection term indicates that most types of system in their study had a net export of kinetic energy from North America.

2.5.2 Diurnal Peaks

Both the generation and dissipation of kinetic energy are greater at 0000 than at 1200 UTC, the bias being especially pronounced during summer (Kung, 1967, 1969). This is coincidental with, although

necessarily related to, the diurnal and annual peaks in convective clouds and storms. The peaks may well be related to energy transfers in both directions, downscale and upscale, between synoptic processes and convection.

2.5.3 Meridional Peaks

Both the boundary layer and upper tropospheric peaks of kinetic energy generation and dissipation were maximum near 35N and 52N latitudes on the average (Kung, 1970). This appears to be related to the mean positions of the jet streams. But the latitudes also happen to coincide with the NSSL (Oklahoma) and AHP (Alberta) study regions respectively, long known to have high frequencies of convective storms. The jet streams and the peaks of energy conversion may both be related to these peaks in convective intensity over North America, thus reflecting downscale and feedback interactions.

2.5.4 Important Scale Interaction Zones in the Cyclone Vicinity

Most energy budget studies have made use of area averages in their computations. Thus, it was not possible to distinguish any 'preferred' regions of activity of the individual budget terms within the vicinity of a cyclone or anticyclone. However, some inferences can be made, supported by a number of cyclone case studies as follows.

- a) For closed systems, such as the entire atmosphere, the total conversion of APE to kinetic energy can be represented by the term (CA) in (2.9). However, as pointed out earlier, for limited regions

of the atmosphere, the APE may be converted to KE either within or outside the analysis region. That is, one must consider transport across the boundaries. Eddy (1965) applied a Lagrangian formulation, where the horizontal and vertical fluxes of KE, (HFK) and (VFK) in (2.6) do not appear, to an 'open-system', developing cyclone. By neglecting these boundary terms, however, one can only hope to approximate a closed system; we refer again to Rossby's quote at the beginning of Chapter 1. Eddy found the most intense KE production near 500 mb in the zone ahead of the cyclone during early development.

- b) Kung (1966) computed the Eulerian KE budget for North America, using a 6-month average in order to minimize errors. Kung found the most intense KE generation ahead of cyclones, but with consistent peaks at 300 and 900 mb, and, contrary to Eddy's (1965) results, a minimum near 500 mb. Since Eddy omitted the boundary terms, Kung's maxima are probably closer to the truth for the average case, although this does not discount the case study results by Eddy. Their vertically-integrated totals of average energy conversion (APE to KE) agreed within 20%.
- c) In a single case study of a rapidly deepening east coast cyclone, Sechrist and Dutton (1970) showed that a source of kinetic energy associated with a jet maximum northwest of the storm, provided a significant amount of the cyclone's energy at lower levels southwest of its centre. They deduced that this occurred by way of parcels decelerating and descending from the upstream jet. Some of the kinetic energy fed directly into the developing cyclone, while the remainder increased the baroclinicity southwest of the cyclone through meridional displacement of airmasses, and subsequent

intensification of the vertical wind shear. This further caused air parcels to ascend and move northeastward from the barotropic region in the warm sector, which then produced a large supply of kinetic energy to the east (cross-isobaric generation and feedback to the planetary scale). They deduced that this new supply of energy probably became a source for storm development downstream. More recent diagnostic studies have confirmed this theory, including Chen and Bosart (1977), and Alpert (1981).

Analyses such as the ones described above raise the question of whether similar processes occur between the synoptic scale and the severe storm environment. We shall attempt to answer this question in the remainder of this chapter, supported by the analysis results which follow in Chapters 4 through 7.

2.5.5 In Situ Energy Conversions

Bullock and Johnson (1971) and Vincent et al. (1977) computed mean values of "generation of APE through latent heat release" (Term G in (2.9)), of about 10 W m^{-2} . These are comparable to the mean values of "cross-isobaric kinetic energy generation" published by Kung and Baker (1975). Smith (1980), however, estimates that the eddy conversion of potential to kinetic energy for synoptic scale mid-latitude cyclones is some 30 times greater than values of cross-isobaric generation of kinetic energy. His explanation is that a substantial amount of APE is exchanged with the surrounding environment with little impact on the cyclone evolution. In other words, most of the energy appears to 'flow

through' the system. At the sub-synoptic scale, Tsui and Kung (1977) found similar magnitudes for the eddy generation and dissipation terms in various types of convective storm situations.

2.6 KINETIC ENERGY EQUATION FOR EDDY PERTURBATIONS

The NSSL upper air network of 10-11 stations with a mean spacing about 100 km was implemented during 1966-67. This provided for the first time extensive upper air data sets for mesoscale studies. McInnis and Kung (1972) published the first energy budget results at this scale. Kung and Tsui (1975), and Tsui and Kung (1977) followed up the initial study at this scale, using an eddy perturbation formulation. These three studies will now be referred to as MK72, KT75, and TK77 throughout the remainder of this chapter.

KT75 derive the "Eddy Kinetic Energy Equation" from the equation of motion, partitioning variables into mean values for the network area and perturbations from these means, such that the dummy variable

$$X = [\bar{X}] + X^* \quad (2.11)$$

The eddy kinetic energy per unit mass is

$$K_e = \frac{1}{2}(u^{*2} + v^{*2}) \quad (2.12)$$

From this, one can derive the KE equation for the area-mean flow (see KT75), which includes eddy perturbation terms. Subtracting that equation from (2.6) yields a formulation for the Eddy KE Equation for an area:

$$\begin{aligned}
 \left[\frac{\partial K_e}{\partial t} \right] &= \underbrace{\left[-\vec{\nabla} \cdot \vec{\nabla} K_e \right]}_{(\text{HFK}_e)} + \underbrace{\left[-\frac{\partial \omega K_e}{\partial p} \right]}_{(\text{VFK}_e)} + \\
 &\underbrace{\left\{ -\left[u^* \omega^* \right] \frac{\partial [u]}{\partial p} - \left[v^* \omega^* \right] \frac{\partial [v]}{\partial p} \right\}}_{(\text{C})} + \underbrace{\left[-\vec{\nabla}^* \cdot \vec{\nabla} \phi^* \right]}_{(\text{CK}_e)} - \underbrace{[D]}_{(\text{D})} \quad (2.13)
 \end{aligned}$$

where the terms are as follows:

HFK = horizontal transport of eddy KE,

VFK = vertical transport of eddy KE,

C = energy conversion from the area-mean flow to eddy disturbances (internal source of KE),

CK = cross-isobaric conversion of APE to eddy KE due to local ageostrophic acceleration of the flow, and

D = dissipation of eddy KE to other scales of motion.

By applying (2.8) to the NSSL data, sub-synoptic scale features may be revealed; i.e., the analysis may present features not observed with synoptic scale observations alone. However, one must realize that the larger scale flow is not filtered out by using (2.8). In the absence of a significant sub-synoptic pattern then, (2.8) will largely reflect the prevailing synoptic circulation. By applying (2.13) to the same data, one observes sub-synoptic scale (eddy KE) features only; i.e., the synoptic scale circulation is subtracted out. This does not claim to adequately account for the kinetic energy balance for features smaller than those detected by the observational network.

2.7 SUB-SYNOPTIC SCALE DIAGNOSTIC STUDIES OF ENERGETICS

Based on the above three studies, four key observations on energetics as related to convection at the sub-synoptic scale are summarized here.

2.7.1 Similarity of Processes at Different Scales

The first may be stated simply as:

Similar energy processes (generation, dissipation) occur at vastly different scales.

The cross-isobaric kinetic energy generation terms of the mean flow have similar vertical profiles, with similar magnitudes at both the synoptic and sub-synoptic scales. This phenomenon is evident in the mean flow profiles of kinetic energy generation in Figure 2.2 (reproduced from MK72).

2.7.2 Scale Interactions with Convective Storms

The second observation is that:

Energy exchanges take place between the synoptic and sub-synoptic scales prior to and during severe convective storms.

Energetics studies for cases of intense convection at the sub-synoptic scale, suggest that kinetic energy dissipation from the synoptic scale west of severe storm complexes, provides kinetic energy (via sub-synoptic scale or mesoscale eddies) to the convective complex (MK72, KT75, and TK77). Also, the upscale conversion of latent heat energy of

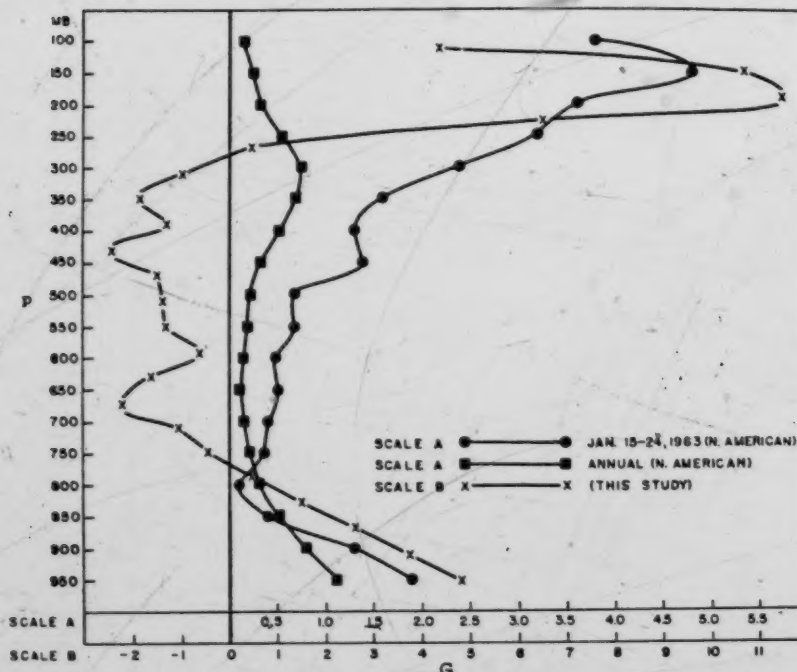


FIGURE 2.2: Comparative vertical profiles of kinetic energy generation ($\text{W m}^{-2}(50 \text{ mb})^{-1}$) for the North American synoptic scale flow: (a) annual mean; (b) 10-day winter average; (c) the NSSL sub-synoptic scale flow 9-day average (after McInnis and Kung, 1972).

convection to KE of the synoptic scale flow appears to take place east of the main convective region. These deductions are similar to those given by Sechrist and Dutton (1970) for the planetary and synoptic scale interactions of a developing cyclone.

However, this energy feedback from severe storms to the synoptic scale may not be all that significant to the larger scale. Using data extracted from MK72, KT75, TK77, and Kung and Smith (1974), Table 2.1 summarizes a synoptic/sub-synoptic comparison of average and case study values of KE generation and dissipation. 'Whole' systems, either an

individual cyclone or a single convective complex are being compared here, even though the areas involved are not comparable.

TABLE 2.1: Comparative values of vertically- and area-integrated totals of kinetic energy generation and dissipation, as computed from Equation (2.8) for average synoptic scale cyclone cases (Kung and Smith, 1974), the average of nine sub-synoptic scale cases (KT75), a single severe convective case study (TK77), and for the mature stage of a severe storm (from MK72).

SCALE	Typical Areas (km ²)	Rate of KE Generation (W m ⁻²)	Approximate Areal Total Generation (TW)	Rate of KE Dissipation (W m ⁻²)	Approximate Areal Total Dissipation (TW)
Synoptic	4,000,000	16.2	65	27.4	110
Sub-Synoptic	40,000	28.3	1	30.2	1
Severe Storm Case Study	40,000	110.1	4	200.4	8
Convective, 1.5-hr interval, Mature Storm	40,000	610.7	24	621.9	25

Scale Lengths: Synoptic, >2000 km; Sub-Synoptic, about 200 km.

While the average rates of KE generation and dissipation for the two scales of motion are comparable (lines 1 and 2), the total area-integrated amounts indicate that the synoptic scale values are about two orders of magnitude larger than at the sub-synoptic scale. Using TK77's most energetically-active case study (line 3), the sub-synoptic area-totals do approach 10% of the average synoptic totals. MK72's maximum area-total values for the mature stage of a severe

convective storm (line 4) were about 25% of the synoptic average, but such values apply only for short intervals, of the order of 1-2 hours.

These numbers suggest that severe convective storms are capable of feeding back a significant supply of KE (primarily through latent heat release) to the synoptic scale, and thereby influencing the synoptic flow, but only for short periods, possibly a few hours at most. An additional significant point here is the fact that the sub-synoptic rates of KE dissipation and generation appear to be about equal. KT75 also note that the rate of mean flow dissipation in the boundary layer for severe storms was about twice that observed in the absence of such storms. This dissipation must be balanced against the energy generated by the storm. Therefore, it is difficult to imagine that one severe storm complex could significantly influence the parent synoptic scale feature for even one synoptic time period (12 hours). The ensemble influence of widespread convection over a synoptic scale region, which might include several such complexes, may be a different thing of course.

Unfortunately, no energetics diagnoses have been conducted to date for the synoptic scale region (continental U.S.) for these same case dates. Such an analysis might reveal the true influence of severe storms on overall synoptic scale features. By computing energy budgets both scales for a number of case studies, it may be possible to determine the conditions under which the synoptic scale contributes significantly to the smaller scale. Ramage (1976) refers to "turbulence bursts" which commence the process, and which he considers non-deterministic. A casual observation by the author from forecasting experience, however, is that convective outbreaks in Alberta seem to

P

occur under the influence of decaying or at least weak synoptic systems, rather than during significant cyclogenesis. With developing synoptic systems, convection tends to be more widespread, without organized severe convective complexes. Since these synoptic periods are deterministic to a large degree, then perhaps the turbulence bursts are equally predictable.

The sub-synoptic scale range links the energy flow between larger (synoptic, planetary) and smaller (meso) scales of motion (Kung and Behrens, 1974). This is illustrated by Figure 2.3 (after KT75), where the vertical profiles of kinetic energy generation and dissipation for the 'mean flow' (Figure 2.3a, computed via (2.8)), are distinctly different from that of the 'eddies' at the sub-synoptic scale (Figure 2.3b, computed via (2.13)). The mean flow profiles have peaks of generation and dissipation near the surface and in the upper troposphere. The "eddy" profiles show the same peaks near the surface, but both terms reverse sign in the upper troposphere. KT75 interpreted this to be eddies adiabatically converting kinetic energy to APE at jet stream levels, primarily as the result of latent heat release. The negative dissipation values represent energy input from other scales of motion, larger or smaller, to the observed scale (TK77). If this does represent energy input, then the term "dissipation" is very much a misnomer.

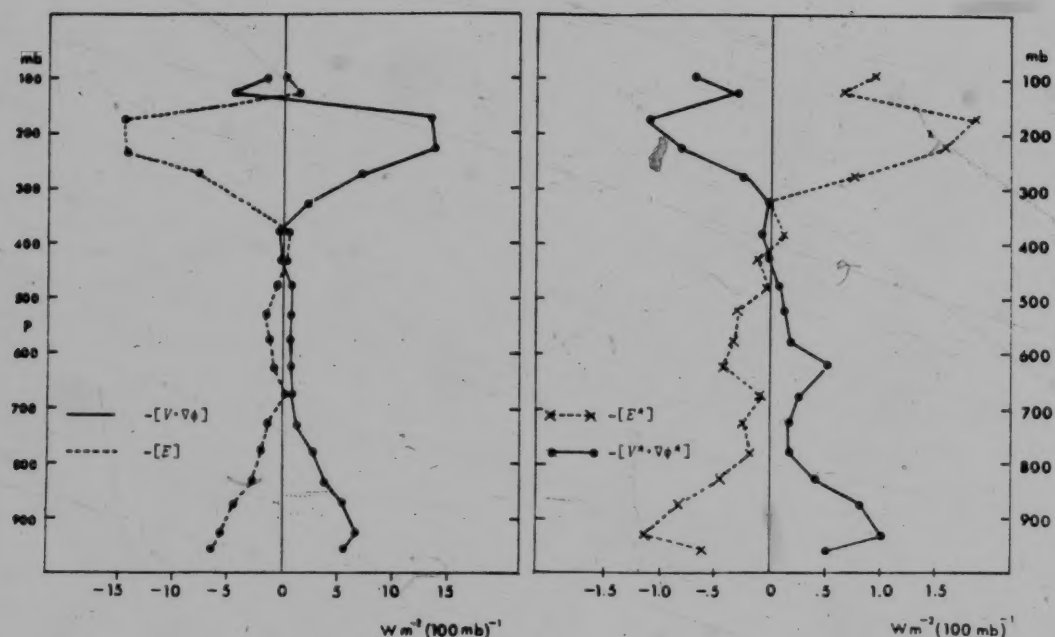


FIGURE 2.3: Mean vertical profiles of kinetic energy generation and dissipation for (a) the mean sub-synoptic scale flow (using (2.8)), and (b) the eddy disturbances at the sub-synoptic scale (using (2.13)). (After Kung and Tsui (1975), using NSSL data.)

2.7.3 Large-Scale Energy Consumption by Thunderstorms

The third observation of sub-synoptic scale energetics is that

Thunderstorm activity consumes kinetic energy supplied from the larger scale in addition to that generated through latent heat release.

While Kung (1966) and Kung and Baker (1975) showed that a significant amount of energy was transported out of a typical synoptic scale cyclone, MK72 revealed that a net inflow of kinetic energy occurred within the (NSSL) sub-synoptic grid area during convective storms.

This is supported by similar studies using coarser resolution data. For example, Fuelberg and Jedlovec (1982) computed a net horizontal KE convergence of 21.3 W m^{-2} for the Red River Valley tornado outbreak of April, 1979. Robertson and Smith (1980) found high values of horizontal convergence of 54.0 W m^{-2} for the Palm Sunday tornado outbreak of April, 1965 (Fujita et al., 1970), and 39.4 W m^{-2} for the Jumbo tornado outbreak of April, 1974. Intuitively, one is inclined to infer that this net KE inflow must be consumed by the sub-synoptic scale (severe storm) processes. Supporting this is the fact that TK77, in their analysis of the eddy KE budget, noted that mid- to high-level KE generation occurred around the storm (environment), while a KE sink (negative generation) occurred within the intense storm region.

However, there have also been studies which appear to contradict the above, including that of Fuelberg and Printy (1983), the only other group to date to use data with spatial and temporal resolution (75 km and 1.5-3 hours) comparable to those of the Kung group (85 km and 1.5 hours). Their analysis for the SESAME-AVE-V (20 May, 1979) severe storm case, revealed large negative values of horizontal flux of KE (-47.4 W m^{-2}) within the small storm scale network. A later analysis (Printy and Fuelberg, 1984), of the larger synoptic scale network data (with 400-km resolution), yielded a horizontal flux of $+0.4 \text{ W m}^{-2}$ for the synoptic domain. This implies a large rate of KE export from out of the small network to the larger synoptic scale flow, where apparently it was consumed or dissipated through frictional processes. Their results will be considered in more detail later, since the current research will focus heavily on the same case study.

2.7.4 Rates of Energy Conversion

The fourth major observation of sub-synoptic scale energetics to be noted here, is that

The adiabatic conversion of eddy kinetic energy to APE in the middle and upper layers, is a maximum during the periods of maximum convective intensity (Tsui and Kung, 1977).

This agrees with other studies which suggest that the severe storm feeds back some of its generated energy to the larger scale flow.

2.7.5 The Importance of Sub-Synoptic Scale Energetics

The preceding four observations barely scratch the surface of this topic. They are based on a very limited number of diagnostic studies of severe storm situations. Only those of the Kung and Fuelberg groups have made use of sub-synoptic scale data. Thus, quantitative comparisons are extremely difficult to make thus far, because of wide variations in the resolution of the available data, data domain sizes, computational differences, and averaging periods. This is especially so when one attempts to compare results from the synoptic and mesoscales. Add to this the fact that there are wide variations in the types and intensities of both synoptic cyclones and convective storms. The proposed STORM-Central network will provide the opportunity for such multi-scale studies of energetics.

A few general statements need to be emphasized concerning the vertically-integrated totals of various KE budget terms, averaged over severe convective storm life cycles. These are: that the total KE

content (K) over the analysis regions have been moderate to high (generally $1.5-4.0 \text{ MJ m}^{-2}$); that the local rate of change of KE is normally positive over the analysis regions; that, on average, there is a net convergence of KE within the storm regions from the outside boundaries, implying contribution of energy to the severe storm from the synoptic environment; that the cross-isobaric conversion of APE to KE is usually significantly positive, presumably a result of latent heat release; and that usually the dissipation is towards sub-grid scales. Of course, even among the limited number of analyses to date, one can find exceptions to all of these observations. Clearly, the importance of these points demands more attention and further upper air measurements at the sub-synoptic scale and mesoscale. The most useful findings in future diagnostic studies involving convective storms, will probably come from storm time-resolved analyses such as MK72, produced, eddy KE budget analyses similar to KT75 and TK77, and multiple scale analyses (synoptic to mesoscale) such as Fuelberg, and Printy (1984) presented.

2.7.6 Conditional Instability of the Second Kind

The mid-latitude interactions between synoptic scale processes and convective complexes being discussed here, should be distinguished from the theory referred to as "conditional instability of the second kind", or CISK. CISK is normally used to describe scale interactions at tropical latitudes, and regards tropical cumulus clouds and synoptic scale motion as cooperatively interacting mechanisms, similar to the extra-tropical interactions being considered here, but with a different

emphasis.

In CISK, while the large-scale disturbance produces the moisture convergence necessary to drive cumulus convection, the cumulus supplies the heat necessary to drive the larger disturbance itself. These interactions seem to explain the development and motion of intense tropical storms and hurricanes. However, the degree of dependence between the two major scales of motion appears to be, for the most part, different from that at mid-latitudes. CISK will therefore not be discussed further in this study. A more detailed treatment of the subject can be found in dynamics texts, such that by Holton (1979), or in Rosenthal and Koss (1968), who first coined the acronym.

2.8 SUMMARY: THE RELEVANCE OF ENERGETICS TO THIS STUDY

The assumption that important physical interactions occur between synoptic scale processes and the severe convective storm environment, is essential to the hypothesis being developed and tested in this thesis. The energetics theory and diagnostic results reviewed here in Chapter 2, are crucial to the understanding of atmospheric processes at all scales, and especially of the interactions between scales. Results learned from large scale energetics are applicable to the mesoscale to a large degree, and at all scales, one is basically dealing with balances of kinetic and potential energy sources and sinks. Thus, if the winds driven by one scale of motion are "flowing through" some phenomenon at a smaller scale, one can assume that energy exchanges and therefore scale interactions are taking place. This must also be true of severe convective storms, unless one takes the improbable stance that such

storms simply float in the atmosphere without any exchange of matter, like a cork on a river.

Starr (1948) and Rossby (1949) both proved that frictional dissipation alone cannot account for the total energy sink of a closed system. This further means that if a system has a kinetic energy source, whether one is dealing with a developing cyclone or an intensifying convective complex, then it must also have an energy sink which is "independent of friction". This fundamental rule alone, implies that the mesoscale convective complex must interact energetically with the larger scale environment, and it partially explains how feedback mechanisms can be maintained.

Lorenz (1955) resolved some of the questions concerning downscale energy transfer, by partitioning both APE and KE budget equations into zonal (mean flow) and eddy components. He made the distinction between frictional and eddy dissipation, emphasizing once again that friction alone does not account for the energy sink. In fact, the diagnostic studies show that both the kinetic energy generation and dissipation terms are often larger at jet stream levels than in the boundary layer where most frictional dissipation must take place. This is true both for synoptic waves (Kung, 1967, 1969), and for convective storms (MK72, KI75, and TK77). The jet stream dissipation is primarily to smaller scale eddies.

Sechrist and Dutton (1970) diagnosed a case where planetary scale energy was an energy source for a developing synoptic disturbance downstream. The convective studies of Kung and associates suggested a similar arrangement between synoptic scale flow and downstream convective storms. More recent convective energetics studies by the Fuelberg group indicated a feedback of energy from the storm to the larger scale flow downstream. They did not investigate downscale effects. Using values of energy terms provided in the literature and summarized in Table 2.1, however, it appears that the feedback effect of convective storms on the larger scale may not be as significant as the downscale effect.

In any event, both dynamic theory and diagnostic results clearly indicate that thunderstorm activity consumes some energy supplied from the larger scale in addition to that generated through latent heat release. This process is probably most important for the pre-storm environment, while the feedback is probably maximum during the period of maximum convective intensity. These facts will be used in developing a hypothesis for convective storm formation in the next chapter.

"The dissipation of zonal kinetic energy by molecular friction is probably very small. The principal "frictional" dissipation of zonal kinetic energy is instead due to small scale turbulent eddies, and it is principally the kinetic energy of these eddies which is dissipated by molecular friction."

....."We must make a distinction between large-scale eddies, which, roughly speaking, are the eddies large enough to appear on synoptic weather maps, and the remaining small-scale eddies."

E.N. Lorenz (1955)

CHAPTER 3.

SCALE INTERACTION HYPOTHESIS and ALBERTA EVIDENCE

3.1 BACKGROUND TO THE HYPOTHESIS

In the two previous chapters we have reviewed a wide field of meteorology from planetary scale dynamics down to cloud micro-structures. Until recently, the only links connecting these scales of motion were made through studies concerned with forecasting convective storms. We now recognize the interactions between these scales, but still do not understand the processes involved. A starting point in developing this understanding is to investigate those severe storms where the relation between the synoptic and mesoscales appears most obvious. Some of the possible dynamical interactions between large scale processes and convective storms were addressed in the literature review. Kinematic links are fairly obvious in the conceptual model presented by Beebe and Bates (1955), and in the mesoscale forecasting capability illustrated in Section 1.3. A scale interaction hypothesis for Alberta hailstorm conditions is now presented, supported to some degree by examples from two severe hailstorm cases. These are used

along with the foregoing review to formulate a conceptual model of convective storms, linking them more closely with synoptic scale processes. The hypothesis will then be tested on severe storm cases from Oklahoma-Texas in Chapters 4 through 7.

3.1:1 Alberta Hailstorm Conditions

The Alberta Research Council (ARC) Hail Project (AHP) summer operations area of approximately 130-km radius is shown in Figure 3.1. The locations of main surface and upper air observing sites are indicated, along with schematic topographic contours which locate the major mountain barrier (Rocky Mountains) over the extreme southwest portion of AHP. This barrier runs from south-southeast to north-northwest through Alberta and northern B.C. The foothills region, 40-80 km wide east of this barrier, is generally thickly forested with patches cleared as grazing pasture. It contains numerous smaller mountain peaks, mostly 10-25 km east of the barrier.

Most severe hailstorms in Alberta occur with a southwest flow aloft preceding a shortwave trough (Strong, 1979, 1982). Such an upper trough usually promotes surface lee cyclogenesis over the foothills of Alberta. A surface low centre will then often form over southern Alberta if the dynamic features are strong enough over and upwind of the region. North of this low, an easterly component of upslope flow develops in the boundary layer, resulting in local moisture convergence over the foothills (see examples in Appendix I, Figure A1.4). These factors, combined with an unstable airmass, lead to severe hailstorms.

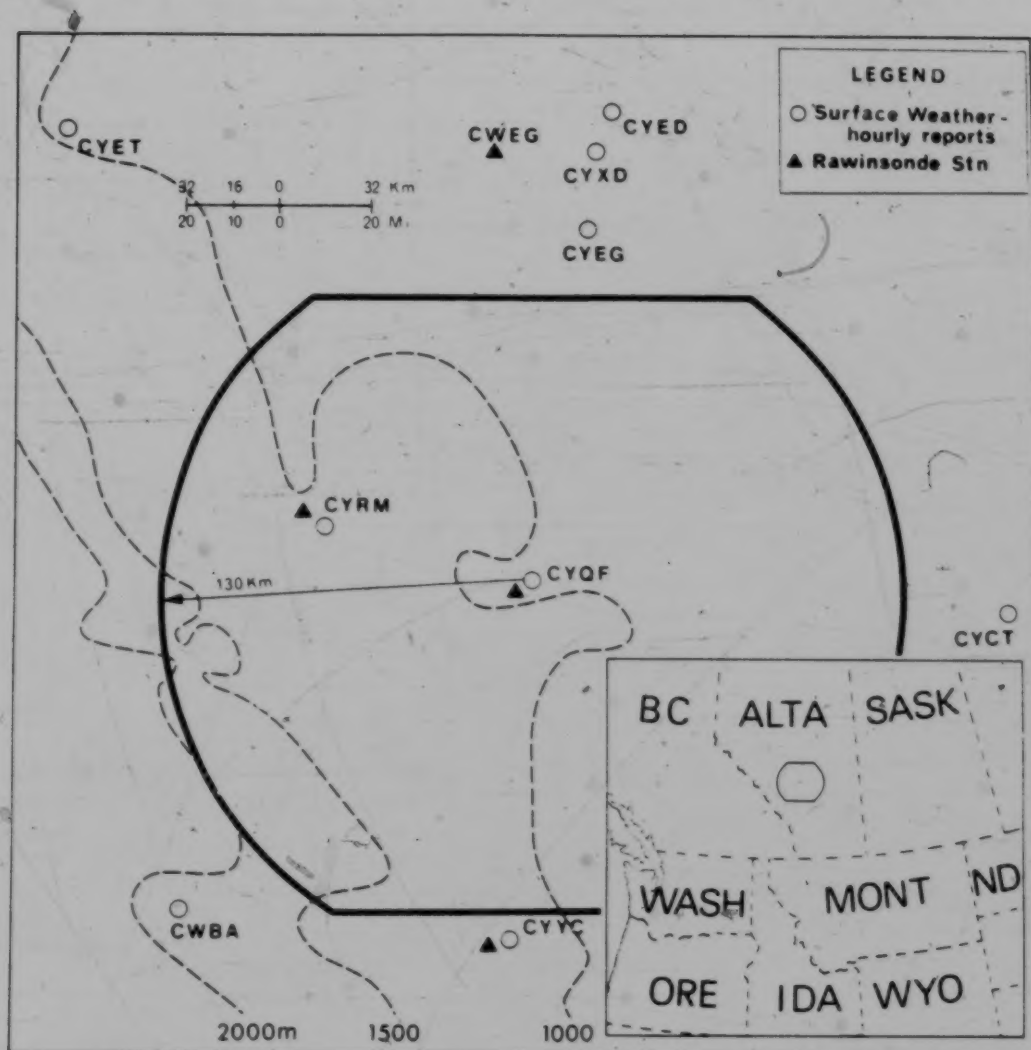


FIGURE 3.1: The Alberta Research Council Hail Project (AHP) summer operations area, showing locations of surface and upper air sites, and schematic topographic contours.

The above type of storm is initiated primarily over the foothills region, and, as observed by the author, often on the east slopes of the foothills peaks. The storm then propagates eastward with the upper flow across mostly grain-producing farmland east of the foothills.

Exceptions to this scenario occur when particularly dry conditions persist over the farmland regions, or in conditions of southeasterly flow at middle levels, when storms either remain over the foothills, or die quickly as they move off. While some surface moisture can be advected from the area south of Alberta on certain days, both the foothills and farmlands also act as local sources of moisture for storms through evapotranspiration. This factor shows up as a clear diurnal trend in surface mixing ratios (Strong, 1979). The major moisture sources in a given situation can control the propagation direction of storms against the mean upper air flow, as was the case for the Calgary and Fort Collins storms discussed in Chapter 1.

3.1.2 The Capping Lid and an Alberta Severe Hailstorm Example

Figure 3.2 shows late afternoon soundings for two consecutive days, 15-16 August, 1973 at Rocky Mountain House (YRM), a region notorious for frequent severe hailstorms. Both soundings indicate distinctive capping inversions around 800 mb, typical of the pre-storm environment for all High Plains regions, and similar to the average pre-tornado sounding of Figure 1.3. The soundings alone provide no real clue as to why convective weather failed to materialize on the first day (15 August), while a particularly severe hailstorm (with hail as large as baseballs) occurred on the second day (16 August)⁷. This storm commenced southwest of the upper air site shortly after the sounding release. The storm-day sounding, in fact, is more stable than that of the first day.

⁷ The hailfall severity of this storm has been documented by Strong and Lozowski (1977), synoptic conditions by Strong (1982).

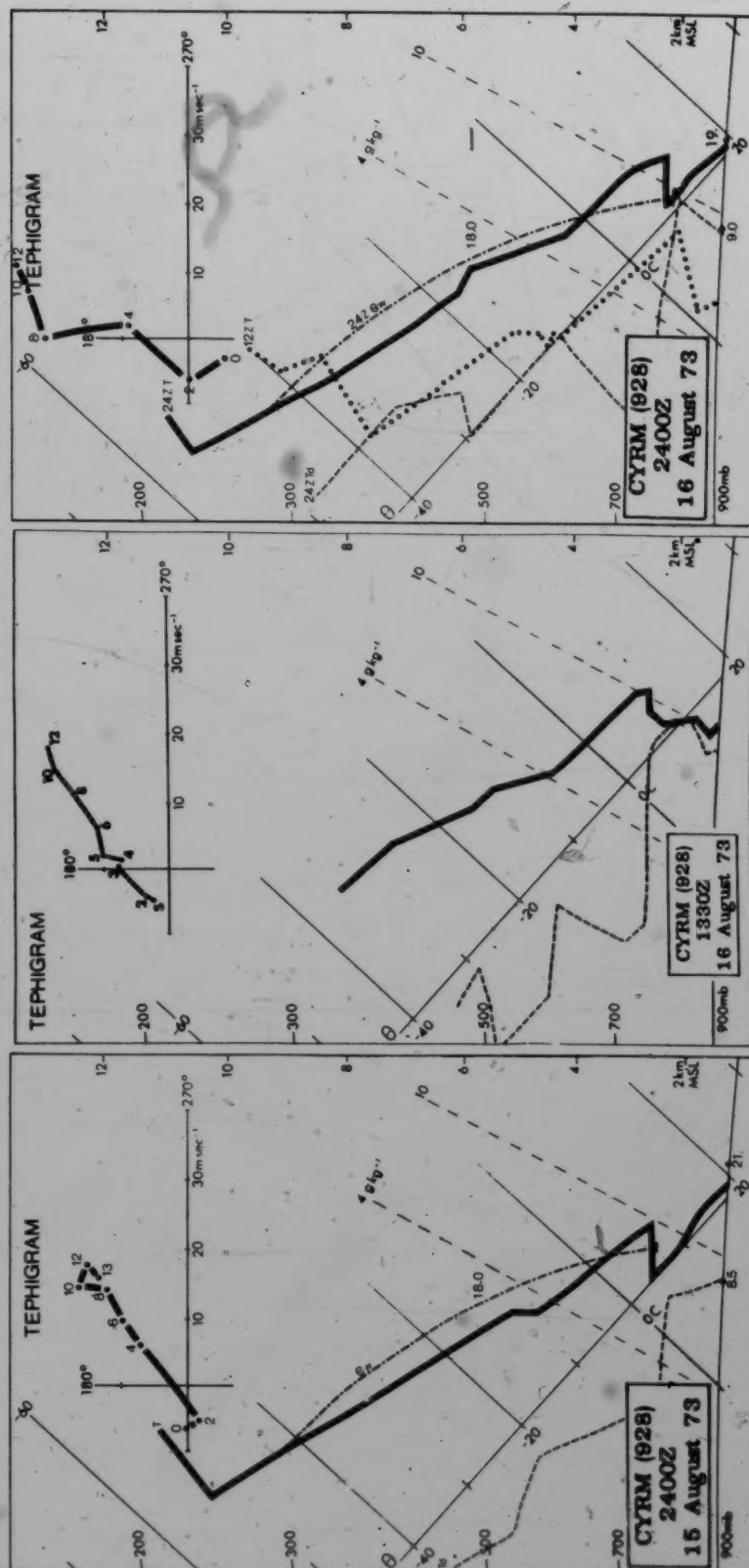


FIGURE 3.2: Representative sounding of dry bulb (solid line) and dew point (broken line) temperatures for 2400 UTC, 15 August, and 1330 and 2400 UTC, 16 August, 1973. The curved line is the estimated cloud adiabatic (Theta-W); the dotted line for 16 August is the post-storm temperature trace at 1200 UTC the following morning.

For this case, it is clear that the capping lid must be at least partially removed at some time prior to storm initiation.

The surface and 500 mb analyses (Figure 3.3) for these two days do show a clear distinction between the two cases at the synoptic scale, as well as clues to the formation and removal of the capping inversion. On the first day there was a shortwave ridge upstream, with associated subsidence over Alberta, a surface high pressure cell, and generally dry air. The ridge had moved by on the second day, and it was replaced by an approaching shortwave trough, strong PVA (for the time of year), large-scale ascent, surface cyclogenesis, and moist surface air.

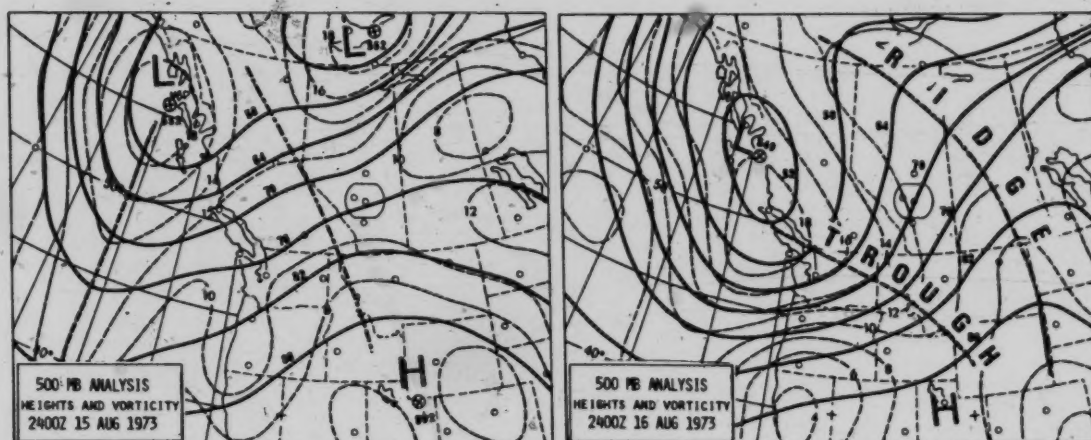


FIGURE 3.3: 500 mb and surface analyses for 2400 UTC, 15-16 August, 1973.

3.2 DESCRIPTION OF THE HYPOTHESIS

A simple hypothesis can now be described which attempts to explain how synoptic scale processes, in combination with orographic flow, prepare the severe storm environment by helping to create and later remove any capping inversion over mesoscale sized regions, thereby

allowing a severe convective storm to take place. The testing of this hypothesis is a major objective of the current study⁸. In the following description, a series of schematic 500 mb and surface charts (Figure 3.4), and vertical cross-sections and soundings (Figure 3.5), are used to illustrate the hypothesis.

3.2.1 Pre-storm Formation and Breakdown of the Capping Lid

The scenario commences with the approach of a synoptic scale (shortwave) ridge, leading to a surface high pressure cell over Alberta, and causing mid-level subsidence (and adiabatic) warming over a broad (synoptic scale) region (Figure 3.4a). Such synoptic scale subsidence is sometimes strong enough by itself to initiate a capping lid at middle levels.

Later, under a southwest flow following the upper ridge passage, the capping lid is enhanced locally near the mountains due to orographic subsidence warming (Figures 3.4b, 3.5a). This enhancement may also commence earlier if the winds are southwesterly in advance of the ridge (such as Figure 3.3 for 15 August, 1973). The author has postulated that, over Alberta, enhancement of the capping lid occurs within a narrow band (perhaps less than 80 km wide) over the foothills parallel to the main barrier. For Alberta conditions, the top of the lid occurs most often near 800 mb, about one km above ground over the foothills. It is broken by high ground, especially in the high valley west of the

⁸ Further tests of this hypothesis under Alberta conditions have now been carried out by the author during the Limestone Mountain Experiment (LIMEX-85), 08-19 July, 1985 (Strong, 1985).

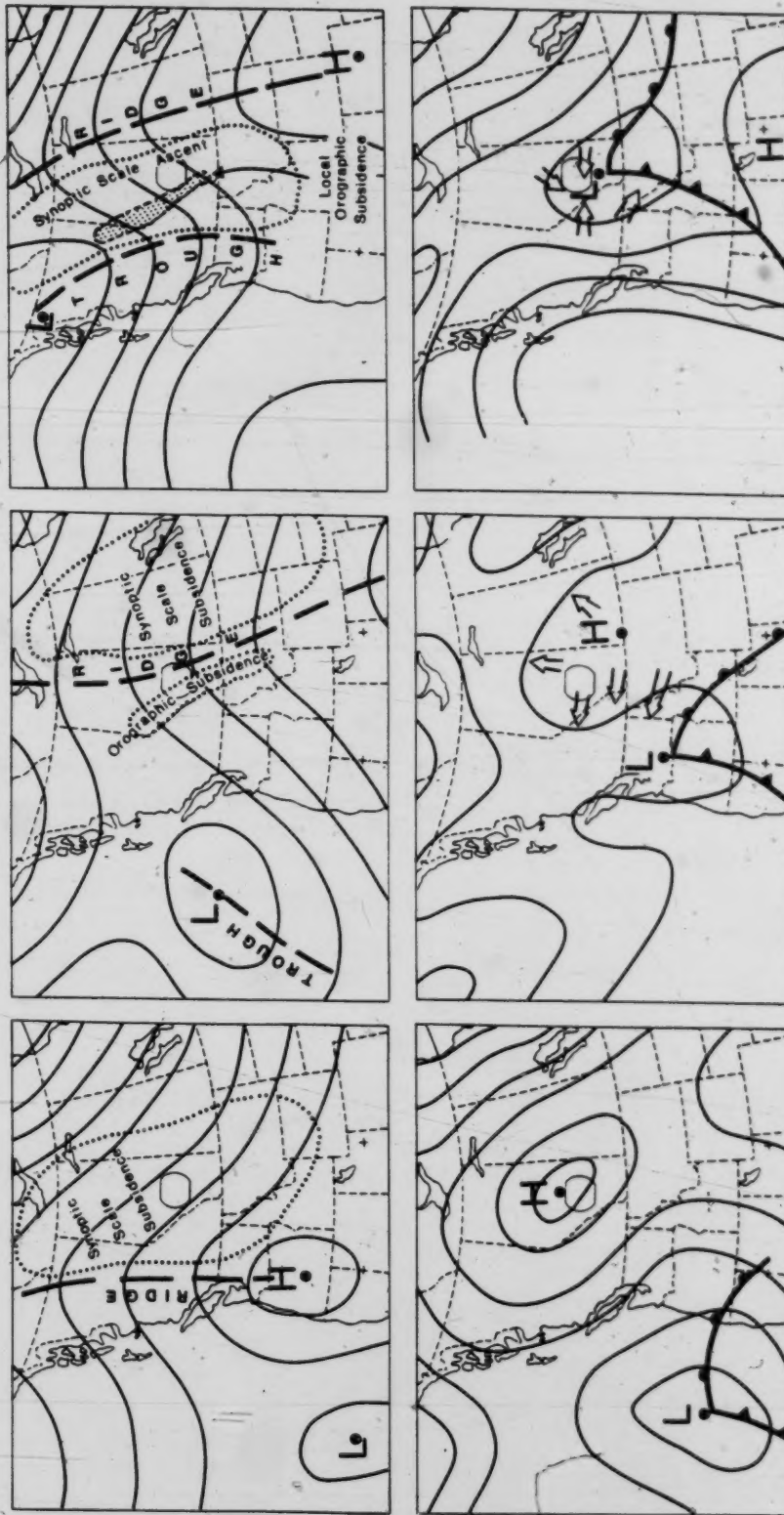


FIGURE 3.4: Schematics of typical summer-time mid-level (500 mb) and surface patterns over Alberta leading to the capping lid formation and severe convective storms; (a) anti-cyclonic flow, initiating the capping lid situation due to associated synoptic scale subsidence warming; (b) post-ridge flow prior to a shortwave trough intrusion, and leading to local enhancement of the capping lid due to orographic subsidence ascent and cyclogenesis over southern Alberta, and initiating the capping lid breakdown and formation of storms over the foothills.

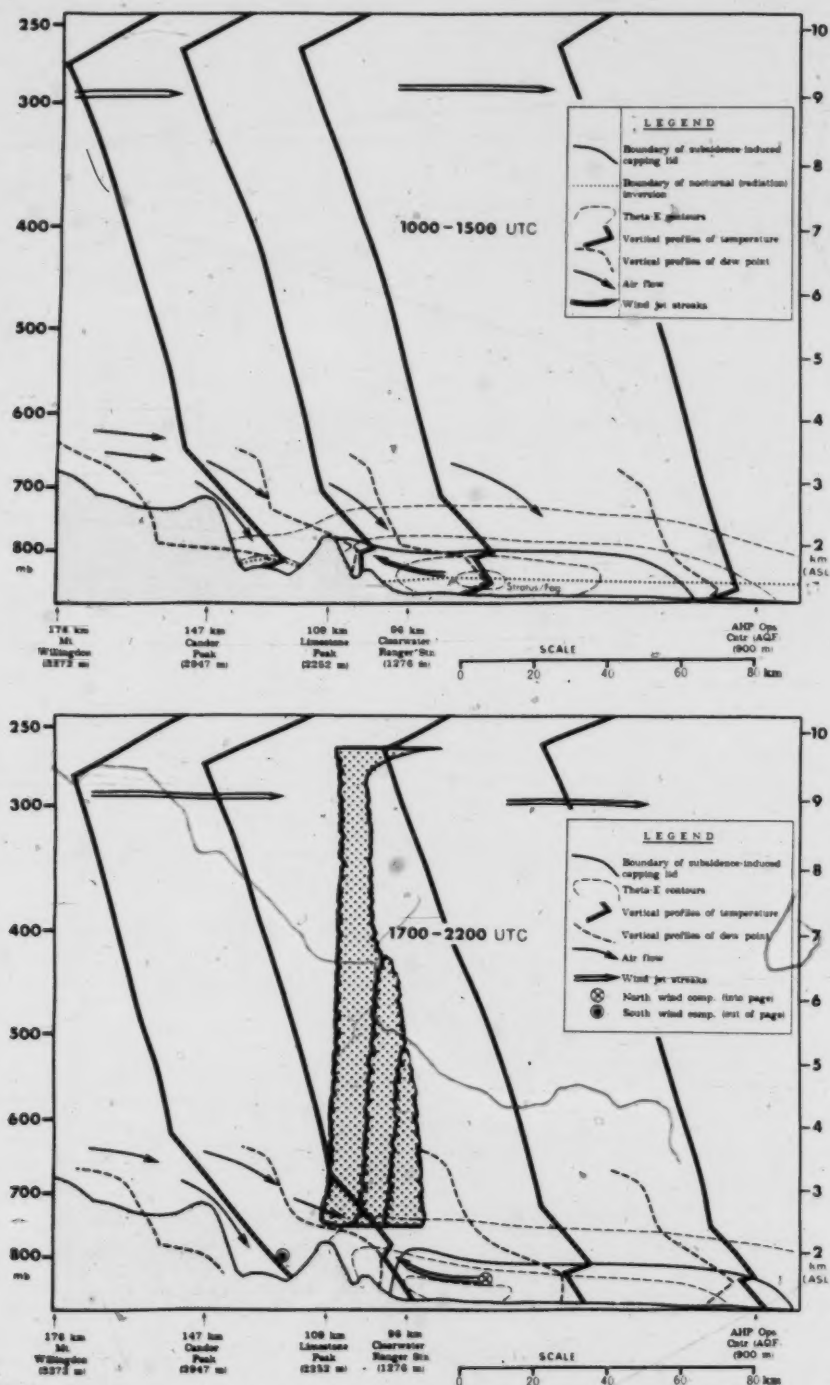


FIGURE 3.5: Vertical cross-sections illustrating the conceptual model of (a) the pre-storm formation and enhancement (1000-1500 UTC) of the capping lid through synoptic scale and orographic subsidence, and (b) the eventual breakdown of the capping lid through synoptic scale and orographic ascent, and formation of a severe convective storm at its western edge by underrunning, ascending moisture (1700-2200 UTC).

foothills peaks noted above, where it is believed that this subsidence extends essentially to near ground level (about 800 mb there)⁹. The vertical cross-section of Figure 3.5a illustrates the combined effects of these processes.

Often, the lid becomes even more pronounced overnight due to surface radiational cooling under clear skies, and a second, lower, nocturnal inversion may form. This combination during summer often results in extensive morning fog and stratus beneath the inversion. Morning observations of such clouds are a clear indication of a well-mixed boundary layer of moisture (surface dew points will probably not dry out), and a readily available source of water vapor for later convective storms.

As an upper trough then approaches with increased southwest flow aloft, lee cyclogenesis¹⁰ over the foothills causes an easterly or southeasterly flow in the boundary layer beneath the inversion, and leads to moisture convergence over the foothills (Figures 3.4c, 3.5b). The increased large-scale and orographic ascent, eventually counteract the orographic component of subsidence, except near the main Rocky Mountain barrier. Adiabatic cooling (and raising) of the inversion then occurs, especially on the east slopes of the most easterly mountain ridges (such as Limestone Mountain or the Brazeau Range in southwest Alberta). Given the right combination of daytime surface heating,

⁹ Some evidence for this is presented in the Limestone Mountain soundings of Figure 3.8, from LIMEX-80.

¹⁰ If the upper flow preceding the trough is from the west or southwest, any surface cold front or trowal involved will normally continue eastward into Saskatchewan. With a southerly flow aloft, the front has been known to stall along the foothills instead, enhancing the instability.

moisture, and latent instability, convective storms then form where the dynamic effects combine to weaken the inversion most, probably most often near the western or northwestern edge of the capping lid region, where localized orographic ascent and increasing synoptic scale ascent can combine to initiate either a convective complex or a line of storms. At this edge of the lid region, boundary layer moisture flows out from under the lid, a process called underrunning (Carlson et al, 1983), thus providing a continual supply of latent heat energy for storms to feed on (Figure 3.5b).

At about this time, as suggested above, large scale ascent has commenced on the west side of the lid region as the upper shortwave approaches. This ascent, possibly in combination with enhanced, localized subsidence outflow downwind from any convective complex now formed, causes the lid to advect eastward. If verified, this would explain the sudden appearance of strong capping inversions at sites downwind from such storms. The author has observed this on occasion on mid-day soundings at Red Deer Airport, where there was no hint of one several hours before.

Meanwhile, warming and drying at mid- to low-levels continue over a narrow localized region close to the mountains and west of the lid, where the strongest component of subsidence persists under the southwest flow (Figure 3.4c). The effect of this warm, dry zone near the mountains west of a region of moisture convergence, creates a low-level mesoscale baroclinic zone, sometimes misinterpreted as a cold front. However, this zone may play an important role in the overall dynamics of the situation, not unlike that of a cold front. It differs from a true front, in that it survives only the lifetime of the convective storm, or

more correctly, until the supply of moisture is depleted by convective storms to the east.

The thermodynamics of the inversion may also create or enhance a low-level jet streak along the rearward edge of the capping lid (Figure 3.5). Such a wind maximum can serve to channel boundary layer moisture into a smaller region of convergence as it underruns the lid. Local orographic ascent then plays a more important role, and a more intense convective complex, as opposed to a line of storms, is apt to form at the northwest edge of the lid region. In this conceptual model of the capping lid, and in the case of either a storm complex or a line, synoptic processes and their interactions with the topography, play the major roles in determining the location, size, and shape of a convective storm.

3.2.2 Other Factors

The orographic component of subsidence is believed to be strong enough to survive both the upper ridge passage, and the effects of a cold frontal passage during the early stages. In effect, the front and associated cloud band pass over the capping lid, leaving the low-level moisture behind, trapped under the inversion. This leads to the more common post-cold-frontal type of severe storm in Alberta, as discussed by Strong (1981), and it produces the climatological pattern of cold (850 mb) advection for such cases as demonstrated in Figure 1.5 (from Longley and Thompson, 1965).

In some areas, hot, dry air mixed over an elevated plateau, is believed to be another major factor in the creation of the capping inversion (Carlson et al, 1983). This may be the case for southern Texas, with a southerly flow from the high Mexican Plateau for example, but it does not appear to be a significant factor in the Alberta situation, with rough, mountainous regions interspersed with deep valleys to the southwest.

3.2.3 Hypothesis Summary

The above stages help to explain the frequent line orientation of thunderstorm complexes in Alberta and elsewhere; i.e., lines of storms parallel to the back (western) edge of a capping lid region. But this does not fully explain the observed occurrence of single quasi-elliptical convective complexes occurring alone in many instances, since synoptic scale processes have been used in the description. However, by including the concept of eddy dissipation, as discussed by Rossby (1949) and Lorenz (1955), the hypothesis can be extended to encompass both convective complex types. It can be summarized as follows:

It is proposed that a transfer of kinetic energy from the larger scales of motion provides (eddy) kinetic energy to the sub-synoptic (or even meso) scale over preferred regions, causing vertical circulations of a similar scale. This leads to (horizontally) differential lifting of the air mass, often in combination with local orographic lifting. The lifting results in adiabatic cooling (of the capping inversion), and subsequent destabilization of this lower layer over 'preferred regions'.

This hypothesis is really an extension of the Beebe and Bates (1955) model. Their model did not receive wide acceptance outside of the forecasting field. It appears that most scientists thought that the Beebe and Bates model ignored the theory that any convective 'trigger' mechanism has to be a mesoscale process, by virtue of the fact that storms form within a region of mesoscale moisture convergence. This feeling apparently is still prevalent (Sanders, 1983). To state the perceived problem in another way: if synoptic scale ascent causing adiabatic cooling leads to de-stabilization of an airmass, particularly one exhibiting the capping inversion, then why do we never see convective clouds, or at least clusters (complexes) of convective clouds with horizontal dimensions of the synoptic scale vertical motion field? The concept of scale interactions and transfer of energy to smaller eddy sizes, makes the above extension of the Beebe and Bates model plausible.

In fact, their model of upper and lower jet structures virtually demands a smaller than synoptic scale region of "eddy" dynamics and ascent (see the hatched regions of Figure 1.3). The area of these regions, which represents the intersection between the areas of divergence and convergence associated with the upper and lower jets, should be proportional to the area of the region of convection. It may be possible to present a more quantitative estimate of this area by determining the hypothesized eddy sizes from perturbation energetics, first presented by Lorenz (1955), and applied to a diagnostic study of convective storms by Kung and Tsui (1975). Thus, under the new hypothesis stated above, it may be possible to predict the approximate size, distribution, and location of the eddies. The eddies would determine the preferred regions for convective formations, and possibly

the actual size of a convective complex. Other factors, particularly topography and net terrestrial radiation, will greatly influence the eddies as well as the synoptic processes involved, making the mathematics of the problem fairly complex. However, unlike turbulence, these two latter effects are fixed and should be predictable from the physics. Mesoscale models of the types described by Pielke (1984), nested within larger scale models to provide boundary energy transfer, may soon have these capabilities.

In support of the hypothesis, a tephigram estimate of the amount of thermodynamic energy required to remove the stable capping inversion layer in one hour, for severe storm soundings such as that of Figure 3.2, is $50\text{--}100 \text{ W m}^{-2}$. This estimate assumes 20% convective cloud coverage over the AHP operations area. Case studies of sub-synoptic scale budgets (by Kung and associates), yield mean-flow dissipation values for the planetary boundary layer of severe storm cases of the order of 100 W m^{-2} . As well, the diagnostic studies of McInnis and Kung (1972), Kung and Tsui (1975), Tsui and Kung (1977), Fuelberg and Scoggins (1978, 1980), and others, reveal peaks of both kinetic energy generation and dissipation at 50-100 mb above the surface (e.g., see Figure 2.3), about the level of capping inversions (Figures 1.3, 3.2). It is possible that eddy dissipation from the larger-scale actually assists surface convection, by first removing the lower stable layer through lift and adiabatic cooling.

Thus, if the amount of kinetic energy transfer and the eddy sizes are small, and if the transfer is primarily to the cloud scale or smaller, then no large storm complex results. This might occur, for example, over regions of surface divergence (high pressure) in advance

of an upper ridge, where there is usually a deficit of kinetic energy. That is, the small amount of energy available is dispersed over a very wide area. On the other hand, with the larger amounts of energy transfer that are diagnosed to occur in cyclonic systems, energy quite possibly is dispersed through larger scale (sub-synoptic to mesoscale) eddies. This leads to the preferred regions of ascent which help to initiate the severe convective complex. The result of this is a well-organized storm, often encompassed by generally clear air because of similar scale regions of subsidence. Carrying this line of reasoning to a smaller scale might help explain why, in any convective complex (storm cluster), the severe storm development is limited to a chosen few cumulus cells.

The nature of the energy transfer may be partially explained from the energetics studies of Kung and his co-workers. It remains, first, to verify the existence of low-level cooling on soundings such as Figures 1.3 and 3.2, "prior to" large cumulus congestus and cumulonimbus formations, and second, to diagnose the size distributions of convective regions and clouds in relation to the regional extent of the hypothesized dynamics. Ultimately, these should be studied in conjunction with an energy budget analysis over the synoptic to mesoscale range. However, the latter goes beyond the scope of the current study.

3.2.4 Extension of the Hypothesis

The last part of Figure 3.2, for 2400 UTC, 16 August, 1973, includes the temperature sounding 12 hours later, revealing a different, colder air mass. Once commenced, the continued ascent and cooling of the air mass allows severe storms to procreate and propagate with the larger scale features. It may be a logical extension to propose that, in effect,

The continued cooling and storm cell regeneration in the immediate environment of a severe storm complex (cloud cluster), is partially a result of a continued transfer of kinetic energy to this scale of motion (or eddy).

It is not the intention here to suggest that this hypothesis is the only, or even major means to create convective storms in any given situation, and thus deny the requirements for surface heating and low-level moisture. Indeed, these other factors, and the latent heat release resulting from cloud formation, are necessary not only to the initiation of convective storms, but also to the feedback mechanisms which balance the larger scale energy budget.

If verified, this concept might also help to explain nocturnal thunderstorm activity, including those storms which are instigated long after maximum temperature time, and even during darkness. Such storms result in approximately 20% of the hail days in Alberta (Strong, 1979).

3.3 THE LIMESTONE MOUNTAIN EXPERIMENT

The first Limestone Mountain Experiment (LIMEX-80) was conducted on 15-17 July, 1980. Its goal was to measure thermodynamic responses to orographic subsidence near the mountains of southwest Alberta, and to determine any effects on the pre-storm environment. A mobile radiosonde system was deployed to the west side of Limestone Mountain (a site designated LMW) in a 10-km wide valley next to the main Rocky Mountains barrier. Its primary purpose was to measure the maximum response to orographic subsidence. Radiosondes were also operated at Rocky Mountain House (YRM) 65 km northeast, and at Red Deer Airport (AQF, AHP operations base), some 110 km east-northeast, in order to detect the creation and breakdown of any capping inversion. These sites and the approximate topography of the region are indicated in Figure 3.6. Soundings were released at three-hour intervals on each of the three days, commencing at 1400 UTC.

A westerly flow aloft persisted throughout the three days as a cold low moved eastward across northern Alberta (Figure 3.7). The average (of four) daily soundings at LMW and YRM (Figure 3.8) shows a strong subsidence warming near LMW, extending virtually to the ground (about 800 mb there), with temperatures 2-8 °C warmer between the surface and 600 mb than at YRM. A severe hailstorm formed west of YRM (over the Brazeau mountains) near noon on 16 July. NOAA-6 visible and infrared satellite images of the resulting convective complex are shown in Figure 3.9, while Figure 3.10 provides a sequence of Red Deer radar PPIs during the mature storm stage. While radar and satellite data indicate that

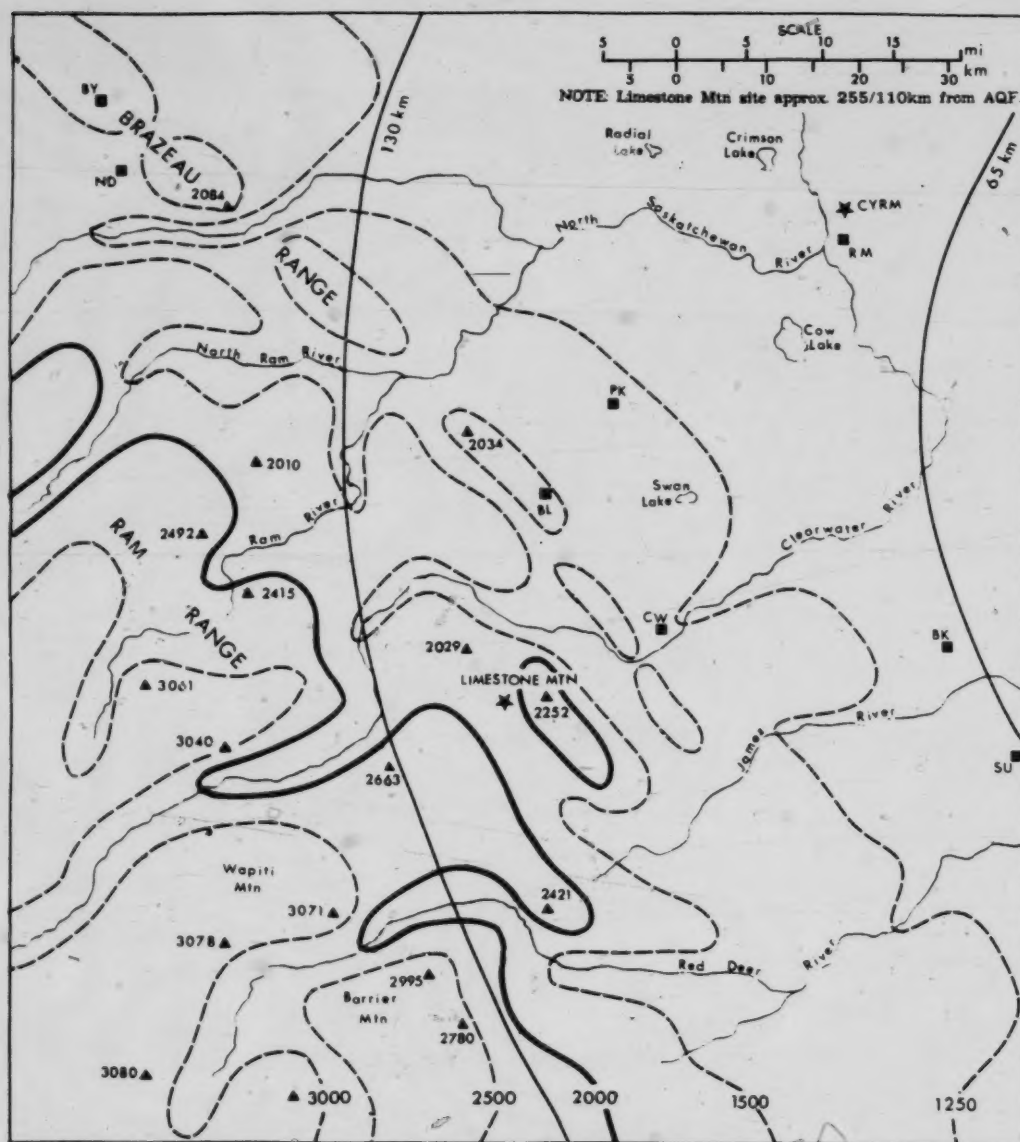


FIGURE 3.6: Topographic chart showing Limestone Mountain and adjacent regions, and upper air sites during LIMEX-80, 15-17 July, 1980.

the convective complex moved eastward, the major hail-producing storm cell(s) propagated southeastward, according to the hailswath analysis (Figure 3.11).

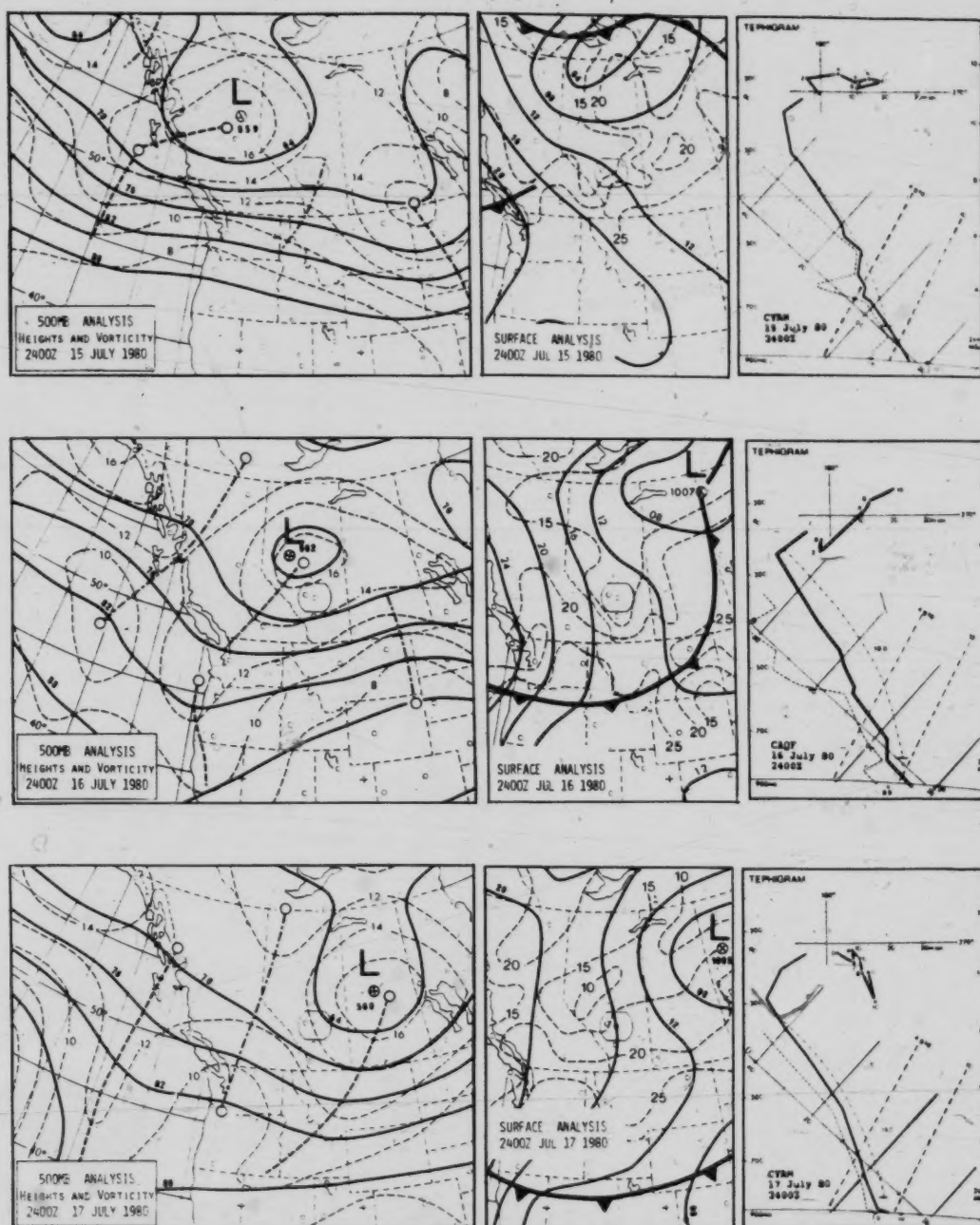


FIGURE 3.7: 500 mb (heights, vorticity) and surface (isobars, isotherms) analyses, and most unstable soundings for 2400 UTC, 15-17 July, 1980 (LIMEX-80). Soundings show temperature, dew point, and maximum cloud pseudo-adiabat (Theta-w); hodograph elevations are in km.

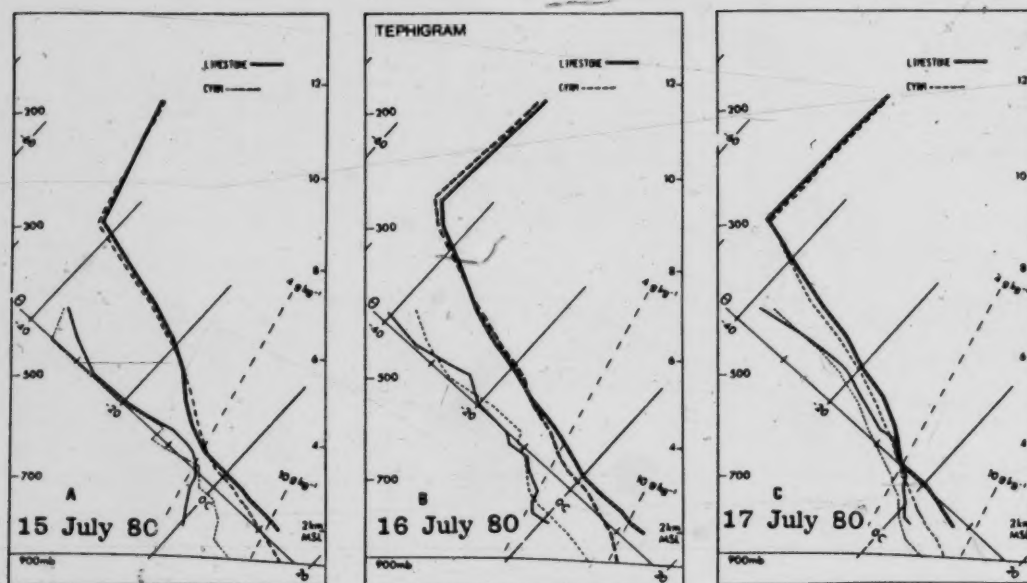


FIGURE 3.8: Average daily soundings (of four each) for Rocky Mountain House and Limestone Mountain for 15-17 July, 1980 (LIMEX-80).

The sequence of YRM soundings for this date (Figure 3.12) reveal a capping lid at 1400 UTC, which was not evident on subsequent soundings. The LMW soundings were, by design, too close to the main barrier and too high (800 mb) to detect the lid. However, it is believed that the lid persisted south and east of YRM for some time. While satellite data is not yet capable of providing the detailed information required to quantify a capping inversion, the subsidence signature can nevertheless be identified in satellite infrared data. Figure 3.13 shows contoured infrared temperature data derived from the GOES-West VISSR satellite data. This data had spatial resolution of 4 km west-east and 8 km south-north over Alberta. (Details of the GOES analysis have been provided by Strong et al., 1984). The warm, dry subsidence tongue near

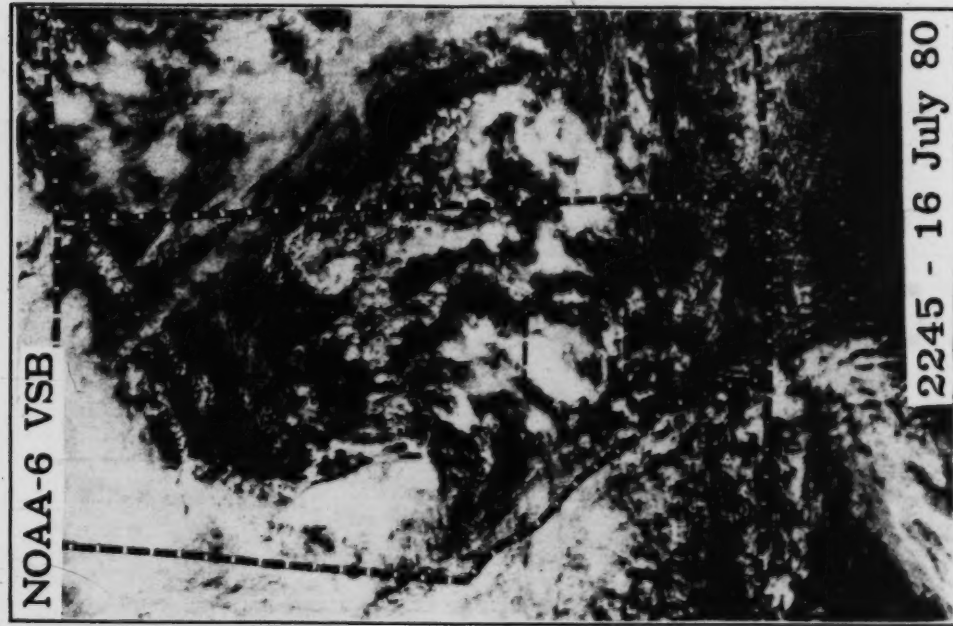


FIGURE 3.9: Visible and infrared NOAA-6 satellite images at storm maturity, 2245 UTC, 16 July, 1980, during LIMEX-80. The convective complex covers most of the northwest quadrant of AHP.

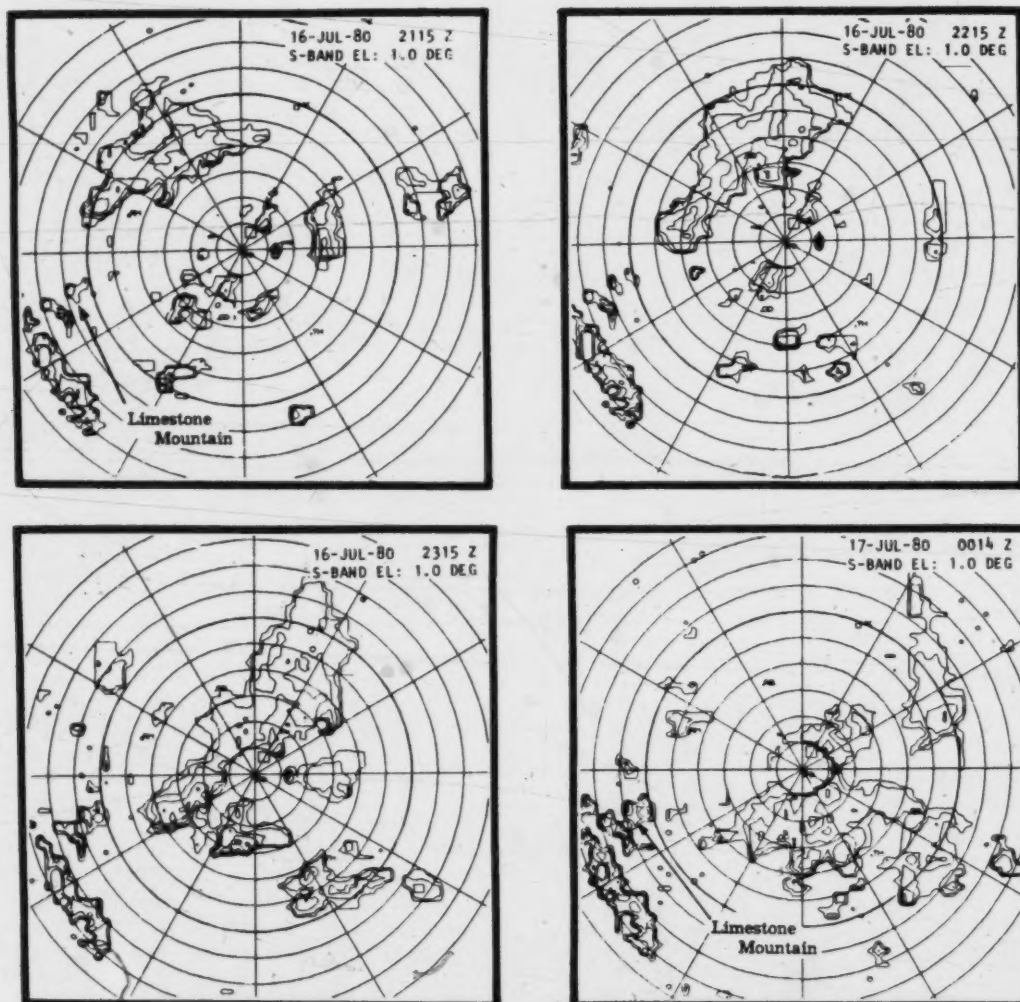


FIGURE 3.10: Hourly sequence of radar PPIs during the mature storm stage, 2115-0014 UTC, 16-17 July, 1980 (LIMEX-80).

Limestone Mountain and southward, shows up as a band some 30-60 km wide. The capping inversion, removed from YRM soundings by this time, is suspected to lie just east of this "summer Chinook" phenomenon, and south of the storms in the vicinity of YRM. A more detailed analysis with additional sounding sites would be required to verify this feature.

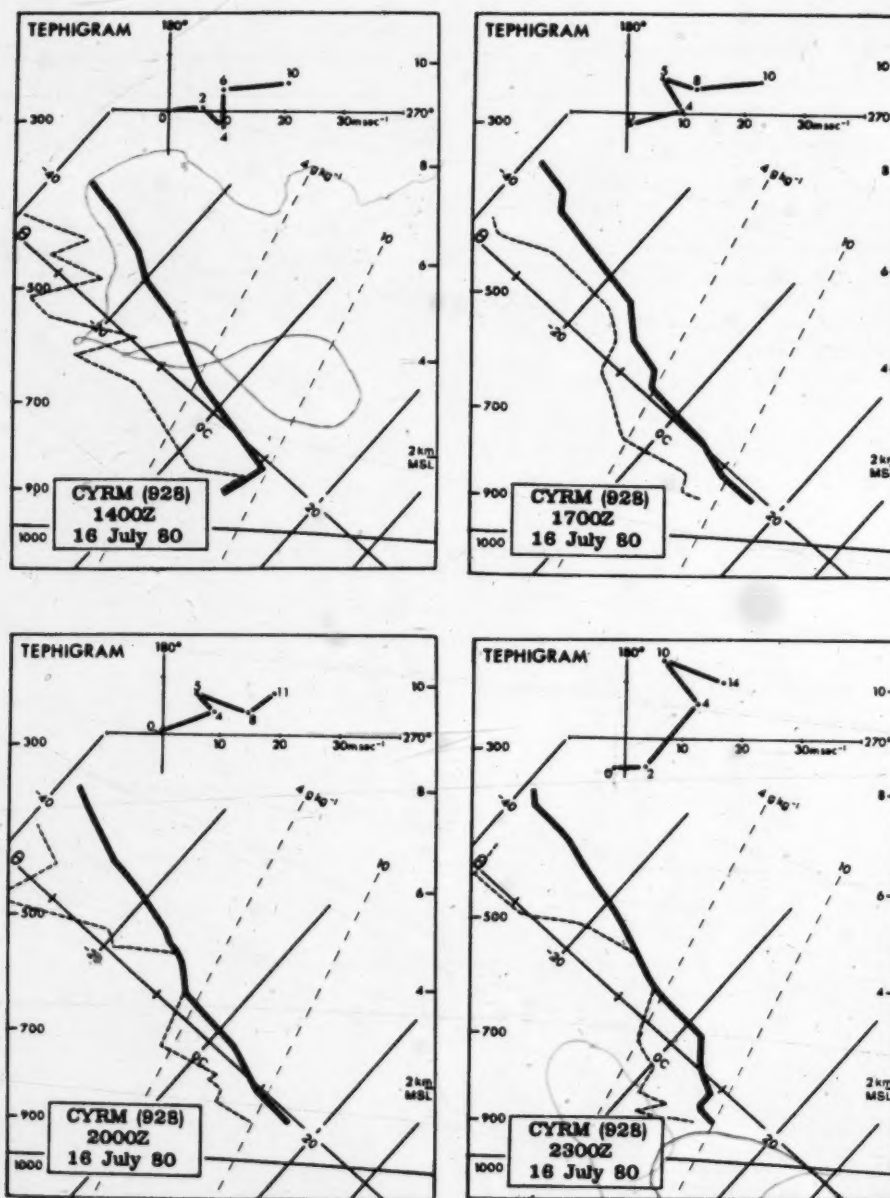


FIGURE 3.12: Time series of soundings at Rocky Mountain House, 16 July, 1980 (LIMEX-80).

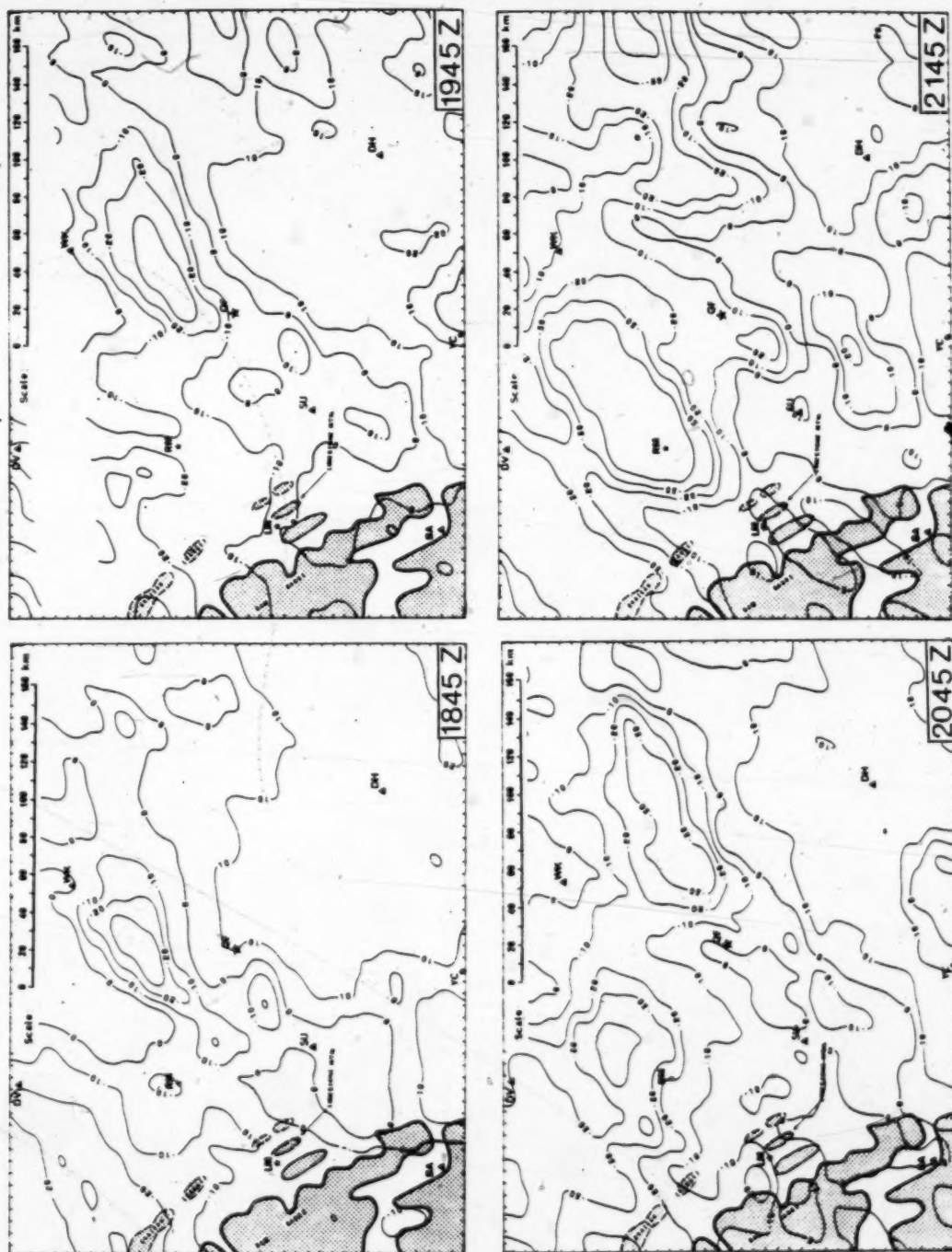


FIGURE 3.13:

- see caption next page -

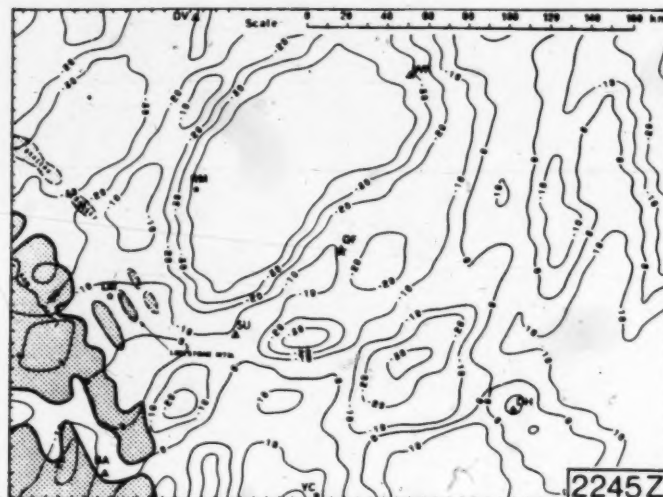


FIGURE 3.13: Contoured infrared temperatures (10°C increments) from GOES-West satellite data, 16 July, 1980 (LIMEX-80), and topography over southwest quadrant of AHP. The four rawinsonde sites (YC, QF, RM, LM) are indicated by stars; stippled regions are mountains;

Analysis of vertical motion fields and energy budget terms would help quantify and test for the hypothesized pre-storm eddies. It is hoped that the LIMEX-85 data will provide some of the necessary evidence.

A one-day test of mobile sounding equipment was conducted on 19 July, 1984, in preparation for LIMEX-85. This day fortuitously turned out to be a moderate storm day with a marked capping inversion at several sites. Aerial mapping of the inversion was also obtained with the Intera/ARC research aircraft facility. These data have not yet been processed, but they may provide more definitive information on the horizontal extent of the capping inversion for Alberta storm cases. Meanwhile, more substantial mapping of capping inversions for Oklahoma/Texas storms was possible using the SESAME radiosonde data. These analyses will be presented and discussed in succeeding chapters.

3.4 SUMMARY

This chapter has offered an explanation for the creation, breakdown, and physical role of a very well-defined and common severe storm signature, the capping inversion (or lid). The capping lid is a frequent pre-storm thermodynamic condition from Alberta to Texas. However, lack of mesoscale sounding data to map and quantify these features, has prevented the determination of whether the lid is by itself a necessary and sufficient storm condition. The conceptual model offered here is based on known dynamics and the published results of research covering different scales of atmospheric motion, including mesoscale studies and the forecasting experience of the author.

A scale interaction hypothesis has been presented which expands on the combined roles of synoptic processes, the capping lid, surface heating, and topographic forcing in organizing the mesoscale moisture field so as to initiate and maintain a severe convective storm. The hypothesis focusses on the necessity for large synoptic scale eddies to transfer energy downscale into various smaller scale eddies before viscous forces take over. These smaller eddies develop their own circulation fields, including vertical motion and surface convergence, and help create the necessary environment for convective storms. Kinetic energy lost through eddy and frictional dissipation, is replaced by the conversion of sensible and latent heat to APE, and eventually back to kinetic energy at the synoptic scale, but not necessarily at the same location.

Thunderstorms can thus be viewed as one of nature's mechanisms which help correct imbalances in the larger scale kinetic energy budget caused by frictional losses. "Energy feedbacks" occur at various scales, depending, it seems, on where they are needed most in the atmosphere; the generation of baroclinic waves or of severe thunderstorms are two such major mechanisms for maintaining the global energy balance.

While the physics of such scale interactions are very complex and are still poorly understood, the explanation and hypothesis presented here, offers many ways to test various aspects of, and to expand on this hypothesis through smaller mesoscale field research experiments and modelling tests. These experiments involve the better use of existing data sources, including current surface and upper air data networks, satellite, radar, and aircraft data, supplemented by additional surface and upper air instrumentation where possible.

The remainder of this thesis is devoted to the verification of some of the processes hypothesized here. Testing of the main hypothesis is accomplished by using a large mesoscale data set not currently available at Alberta latitudes. In the process, important new questions are addressed which may lead to revision of the hypothesis, and thus contribute to the formulation of a "total" conceptual model of severe, as well as weak convective storms.

"When you can measure what you are speaking about and express it in numbers, you know something about it; but when you cannot measure it, when you cannot express it in numbers, your knowledge is of a meager, unsatisfactory kind."

- Lord Kelvin

CHAPTER 4

ANALYSIS TECHNIQUES

4.1 DATE SOURCES AND ANALYSES

The extensive upper air data required for mesoscale diagnostic studies of severe storms just does not exist for Alberta. Even synoptic resolution is relatively poor, with not a single upper air station in neighbouring Saskatchewan. One of the very few data sets to date which appears to meet requirements for both synoptic and mesoscale analysis, is the SESAME data of 1979, with mesoscale data focussing on severe Oklahoma storms. Differences between Oklahoma and Alberta convective storms are primarily in the amount of available boundary layer moisture and the subsequent intensity of typical storms. Krauss and Marwitz (1984) indicate that the severe convective storms of Oklahoma, Colorado, and Alberta involve very similar microphysical processes. Also, while the annual severe storm periods for these three regions peak at different times of the year (May, June and July respectively), the planetary scale mean flow patterns during corresponding storm periods are qualitatively similar. For example, the monthly mean 700 mb pattern (heights and temperatures) over Alberta during July is remarkably similar to that over Oklahoma during May (see Appendix Figures A1.1, A1.2). There is also ample evidence (e.g., Beebe and Bates, 1955;

Fuelberg and Scoggins, 1978; Maddox, 1983; Bosart, 1984) that storms of Oklahoma and nearby regions occur under synoptic conditions similar to those of Alberta. The main similarity is the approach of an upper shortwave trough (see Appendix I, Figure A1.3).

Therefore, many results from studies using SESAME data, will be applicable to Alberta severe storms. Most differences will be found in the "fixed" meso or smaller scale effects; for example, local surface heating or orographic effects. Also, regions at lower latitudes such as Oklahoma have generally higher thermodynamic potential, because of increased sensible heat and moisture availability in the boundary layer.

4.1.1 Data Types and Analysis Regions

The current diagnostic work involves the following two data sources:

- 1) The limited data set of the Alberta Limestone Mountain Experiment, 15-17 July, 1980 (LIMEX-80), which has been discussed already in Chapter 3. LIMEX-80 was carried out by the author through the field research activities of the Alberta Research Council Hail Project (AHP). The analyses include GOES satellite data, as reported by Strong et al. (1984).
- 2) Extensive Oklahoma/Texas SESAME data sets for the Area Variability Experiments (AVE), including (a) AVE-V (20-21 May, 1979), and (b) AVE-IV (09-10 May, 1979), provided by NASA, with various other supporting data from other sources as listed in the SESAME 1979 Data User's Guide (Barnes, 1981).

Analyses of the SESAME cases are based on synoptic upper air data (resolution over south central U.S.: 370 km with 3-hour soundings), surface data (resolution: approximately 130 km and one hour), special SESAME mesoscale upper air network data (resolution over Oklahoma/northern Texas: 20 sites at 80-km spacing and 1.5 to 3.0 hour soundings), GOES-East satellite imagery at one-hour intervals, NOAA radar PPI composite summaries at 1-2 hour intervals, and NSSL WSR-57 radar PPIs on microfilm (one-minute intervals). References to specific convective events were extracted from hourly surface observations, from a severe weather summary provided by July and Turner (1981), and from the NOAA Storm Data summary (1979).

The data types, sources, and spatial and temporal resolution are summarized in Table A2.1 of Appendix II. The topographic analysis of the south-central U.S. region, shown in Figure 4.1, was contoured from site elevations of all surface and upper air stations provided with the SESAME data. The locations of radiosonde stations used in the analyses of this study, including the NWS synoptic sites and special SESAME mesoscale sites, are indicated in Figure 4.2. These stations, their 3-letter identifiers, numbers, elevation, latitude, and longitude, are also summarized in Appendix II (Table A2.2), along with a more detailed map of the SESAME-AVE-IV/V/VI mesoscale upper air network. Figure 4.3 locates most of the NWS synoptic stations used in the surface analyses.

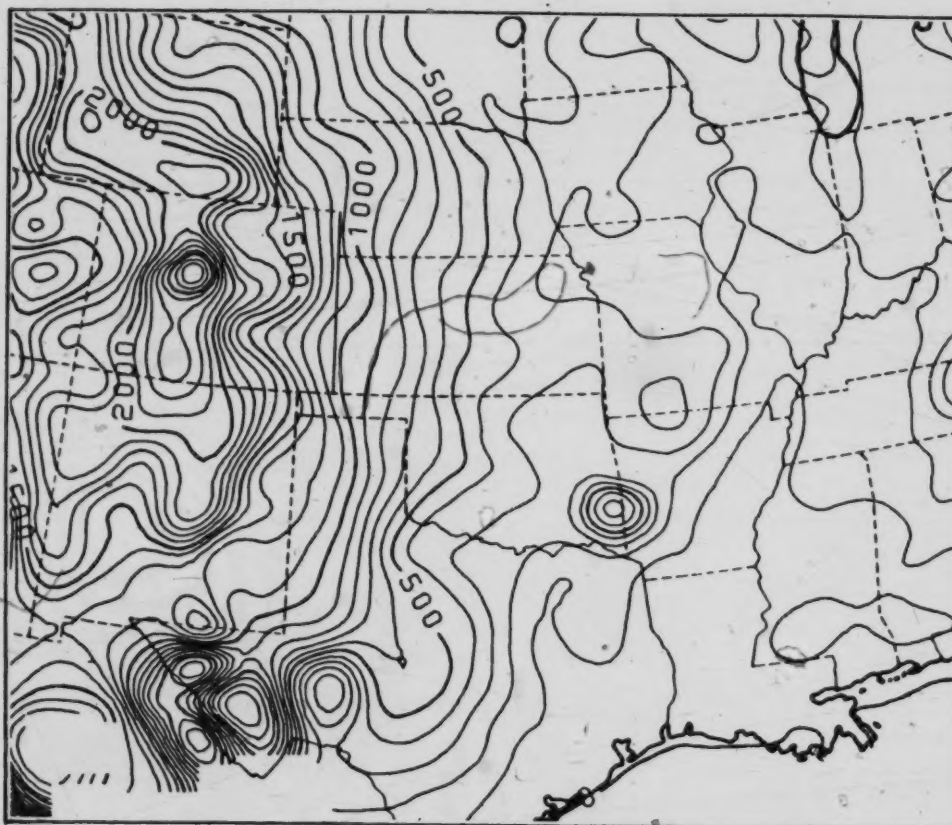


FIGURE 4.1: 100-Meter topographic contours of the central U.S., contoured using the elevations of all synoptic and mesoscale surface and upper air stations.

4.1.2 Types of Analyses

The discussion of results in Chapters 5 and 6 will deal with the SESAME-AVE-V case, and will include the following types of analyses:

- 1) a detailed synoptic scale weather summary, including upper air and surface analyses provided by the Canadian Meteorological Centre (CMC), cloud features from GOES-East satellite imagery, and convective intensity from NOAA radar composite summaries;

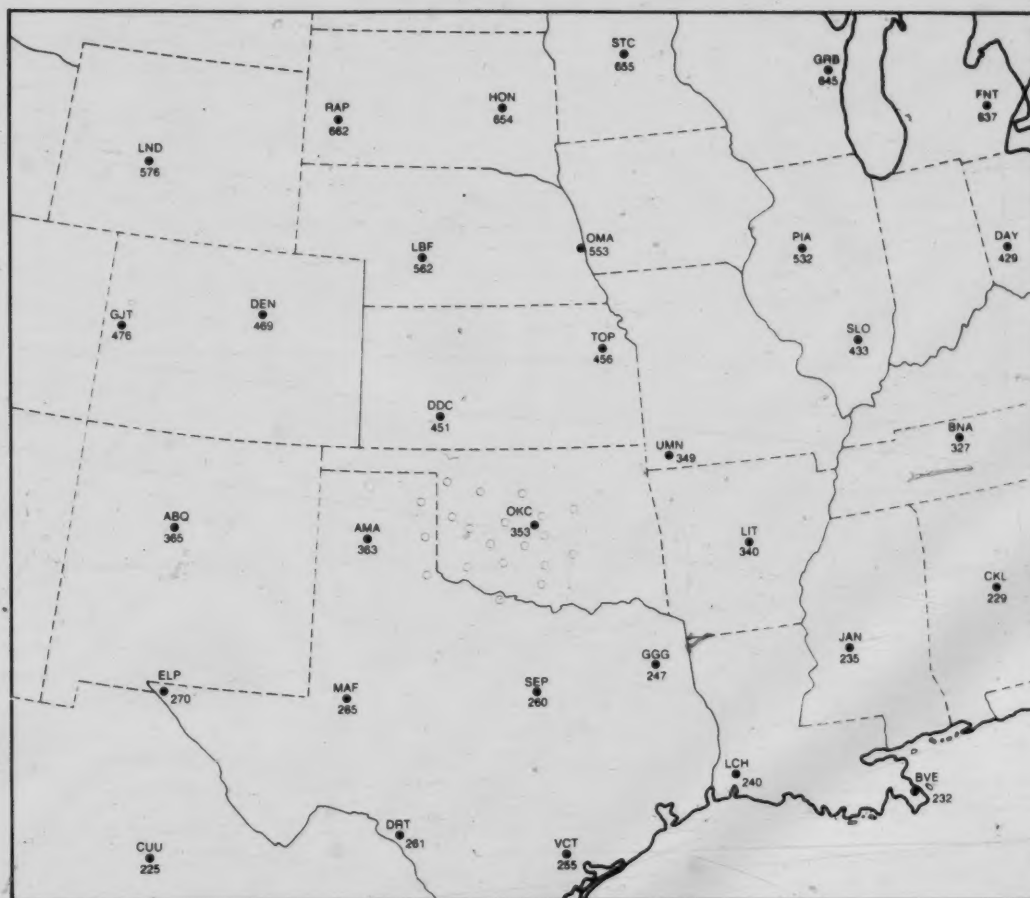


FIGURE 4.2: NWS synoptic (solid circles) and SESAME AVE-IV-V-VI mesoscale (open circles) radiosonde networks used in this study.

- 2) 3-hour synoptic and mesoscale (80-km resolution) isobaric (850-600 mb) analyses of vertical motion fields, with emphasis on the pre-storm environment;
- 3) investigation of the spatial and temporal continuity of observed mesoscale vertical motion features, including vertical cross-sections and time-height cross-sections, and relating these to other mesoscale features;

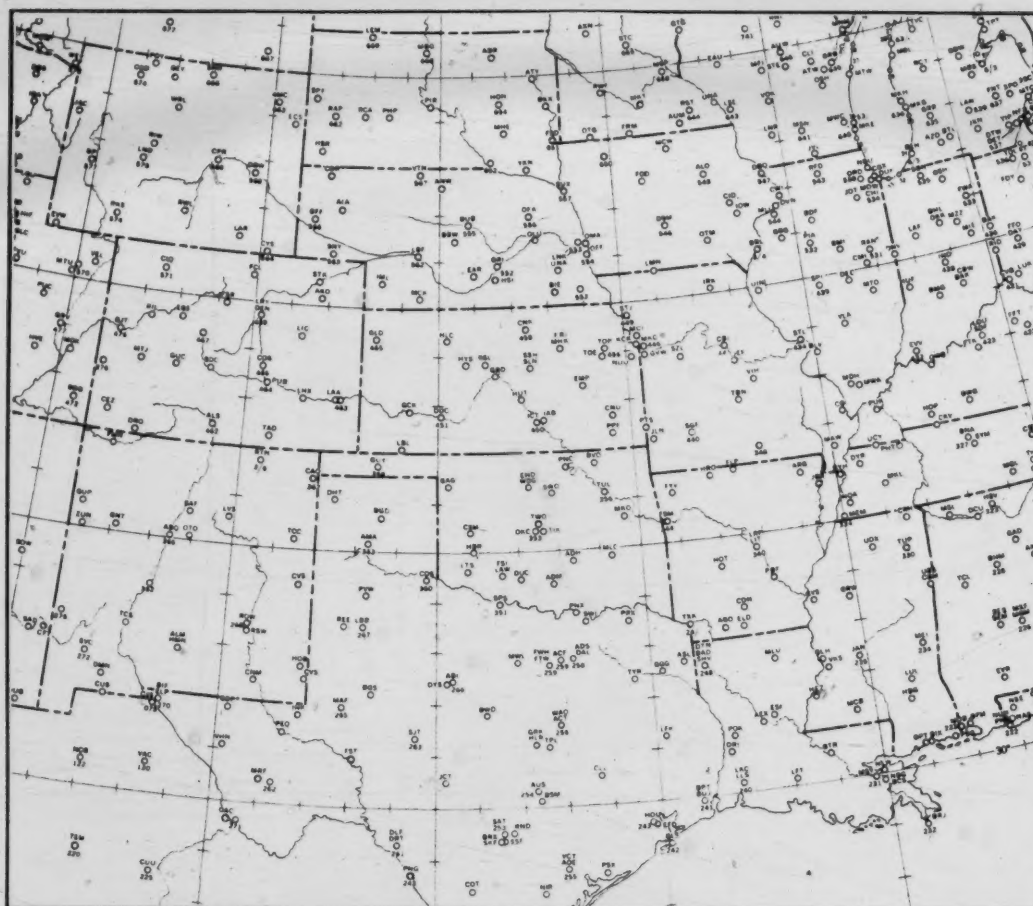


FIGURE 4.3: NWS surface synoptic network.

- 4) mesoscale isobaric analyses of pressure height, temperature, and wind fields;
- 5) hourly sub-synoptic scale (130 km or better resolution) analyses of surface pressure, temperature, and humidity fields;
- 6) detailed analysis of soundings, including qualitative and quantitative comparisons of pre-storm sounding variables as storm prediction tools;
- 7) examination of the capability for predicting the observed convective

complexes, using pre-storm data to compute the forecast Synoptic Index of Convection (Sc4), supplemented by surface Theta-E analyses;
8) verification of the Sc4 using independent data, including the vertical motion analyses, satellite imagery, observations of severe weather phenomena, and total precipitation;

Contoured fields for these analyses were produced using the Surface-II graphics software of Sampson (1978) and the computing facilities of Alberta Research Council. Inter-comparisons of the various data analyses will be largely qualitative, with the greatest emphasis on the AVE-V case of 20-21 May, 1979. Results are also compared with different analyses by other authors who have studied the AVE-V case. The AVE-IV case of 09-10 May, 1979 will be treated as a confirmatory case. It is important to note that the vertical motion computations in this study do not include any orographic components directly. Mesoscale discussions attempt to flag any suspected errors in vertical motion features due to orographic motion or diabatic heating (particularly at 800 mb), or to latent heat release (particularly at 600 mb). In the discussions of vertical motion, emphasis will be placed on the mid-levels (700 mb).

4.2 MEASUREMENTS OF ATMOSPHERIC VERTICAL MOTION

Vertical motion in the atmosphere is closely associated with regions of horizontal convergence and divergence, and the familiar surface pressure patterns of lows and highs which determine our weather. Most meteorological processes require ascent to provide the necessary

adiabatic cooling for condensation, clouds, and precipitation, and subsidence to provide the regions of clear, drier air. Vertical motion is therefore one of the most crucial variables in both meteorological analysis and numerical models, and it is certainly the critical one in the computation of kinetic energy budgets.

For many applications, especially for synoptic scale processes, one is dealing with vertical motions of only a few centimeters per second. Such motions are far too small to measure by ordinary means, and certainly not with rawinsonde balloons, since balloon ascent rates are usually two orders of magnitude larger. Therefore, this very critical variable must be estimated indirectly. The computational techniques used to determine vertical velocity depend on a combination of wind, pressure, temperature, and humidity measurements, and in some cases, they are even based on precipitation measurements (Bannon, 1948). The major techniques used are known as the adiabatic, kinematic, vorticity, and omega equation methods. These are covered in many texts on dynamics such as that by Holton (1979). The latter three methods depend heavily on accurate wind estimates. The omega equation is widely used in numerical models, while the kinematic method is preferred for diagnostic studies, including most of those on energetics mentioned thus far.

4.2.1 Literature Comparisons of Vertical Velocity Computations

Intercomparisons of the kinematic, omega equation, and adiabatic methods have largely been limited to synoptic scale data. In these cases, only order of magnitude agreements has been obtained. Smith (1971), using four days of synoptic data, could not declare one or the

other of the kinematic or omega methods superior. Using a more objective technique for comparing these methods with observed cloud and precipitation fields, Smith and Lin (1978) found that the kinematic method gave more realistic synoptic scale vertical motion than either of two versions of the more complicated Omega equation method.

Hansen and Thompson (1965) judged the kinematic and omega methods superior to the adiabatic method for a three-day synoptic case study, although the basis for this judgement is not indisputable. The comparisons were somewhat biased in that the temporal resolution of the available data was 12 hours, while much of the cloud types examined in the verification were convective, and were therefore of the mesoscale. Interestingly, the adiabatic analyses of vertical velocity did show more realistic mesoscale spatial features than either of the other methods (this author's observation). Also, the adiabatic results were improved when the effects of latent heat release were included.

Fuelberg and Lee (1982) provide the only kinematic/adiabatic comparison to date where the computations approach mesoscale resolution. They used the SESAME-AVE-I data, with 3-hour temporal but only 250-km spatial resolution. Employing a statistical verification which uses satellite clouds and radar echoes, they found the adiabatic method equal to, and sometimes superior to the kinematic method at 700 mb, although the kinematic computations held an edge at 500 mb. A qualitative re-examination of their data, however, suggests that the adiabatic regions of ascent corresponded more favourably with convective complexes than did the kinematic method. Deterioration of the adiabatic method at 500 mb appears to stem from errors due to cloud latent heat.

4.3 KINEMATIC AND ADIABATIC COMPUTATIONAL METHODS

The major analysis of this study employs SESAME data and focusses on Oklahoma-Texas storms. However, the author wished to later compare the research results of these storms with those of Alberta, when a reasonable data set (such as the planned LIMEX-85 data) is made available. Therefore, in choosing a technique to compute vertical motion, it was desirable to use one which would be easily adaptable to Alberta data analysis later. Of the techniques mentioned above, the adiabatic method received some early preference, since it does not require so many radiosonde sites as the more commonly used kinematic method in order to compute a field of vertical motion. Both of these methods will now be described.

4.3.1 Kinematic Method

The kinematic equation for vertical velocity is derived by vertically integrating the continuity equation (Holton, 1979), so that

$$\omega(p) = \omega(p_0) - \int_{p_0}^p \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right)_p dp \quad (4.1)$$

where $\omega = \frac{dp}{dt}$ is the vertical p-velocity, p_0 is some reference pressure level, and the integrand is the horizontal divergence integrated over the pressure layer. Because of the divergence term, the kinematic method is dependent on the usually small departures of the wind from geostrophic conditions. As such, it is extremely sensitive to wind errors, so that a 10% error in wind speed can easily cause a 100% error in vertical velocity, especially at the upper boundary. To reduce

this effect, the constraint $\omega = 0$ is normally imposed at the top of the atmosphere, often taken as $p = 100$ mb in diagnostic studies. The lower boundary condition is the surface vertical motion, ω_0 , which is sometimes assumed to be zero. However, ω_0 has contributions due to pressure tendency, cross-isobaric flow, and terrain-induced vertical motion (Holton, 1979). Fuelberg and Jedlovec (1982) found that the first two contributions were always $< 0.3 \mu\text{bs}^{-1}$, but that the orographic component near the Rocky Mountains averaged ± 1 to $\pm 2 \mu\text{b s}^{-1}$.

In order to satisfy these boundary conditions without invalidating ω at mid-levels, various optimization-adjustment schemes have been applied, including those of Kurihara (1961), O'Brien (1970), Kung (1972), and Chien and Smith (1973). However, application of these schemes without modification to suit the nature of the specific wind data being sampled, may not filter errors at all, or may even compound them through aliasing (Pedder, 1981). Some, including Wilson (1976), have used the adiabatic method (see 4.2.2) at 100 mb to adjust their kinematic profiles, instead of setting $\omega = 0$. Fuelberg and Printy (1983) tried this adiabatic adjustment on the SESAME AVE-V case, but they suggested that significant release of latent heat at upper levels during intense convection violated the adiabatic assumption.

4.3.2 Adiabatic Method

The adiabatic equation for unsaturated vertical motion may be derived from the first law of thermodynamics written in the form

$$c_p \frac{dT}{dt} - \alpha \frac{dp}{dt} = \dot{q} \quad (4.2)$$

where $\dot{q}$ represents the rate of heat gain or loss due to diabatic processes such as surface heating from insolation. We shall see later that $\dot{q}$ can be neglected for short periods of time, except in the boundary layer. For now, let us assume $\dot{q} = 0$. Then, by letting $dp/dt = \omega$, and expanding dT/dt , we arrive at

$$\frac{\partial T}{\partial t} + \left[u \frac{\partial T}{\partial x} + v \frac{\partial T}{\partial y} \right] + \omega \frac{\partial T}{\partial p} = \frac{\alpha \omega}{c_p} \quad (4.3)$$

At this point, the usual route in the derivation (e.g., Holton, 1979) is to combine the two right-most terms in (4.3). Then, with the aid of the equation of state and Poisson's equation, one arrives at the adiabatic equation:

$$\omega = S_p^{-1} \left[\frac{\partial T}{\partial t} + u \frac{\partial T}{\partial x} + v \frac{\partial T}{\partial y} \right] \quad (4.4)$$

where S_p , the static stability parameter, is

$$S_p = -\frac{T}{\theta} \frac{\partial \theta}{\partial p} \quad (4.5)$$

or, in terms of lapse rates,

$$S_p = (\Gamma_d - \Gamma) / \rho g \quad (4.6)$$

where $\Gamma = -\partial T / \partial y$, the actual lapse rate, and $\Gamma_d = g/c_p = 9.76$ $^{\circ}\text{C km}^{-1}$, the dry adiabatic lapse rate (Holton, 1979).

4.3.3 Modified Adiabatic Method

One distinct advantage of (4.4) over (4.1) is that it is not nearly so sensitive to wind errors, so that geostrophic wind computations are usually quite adequate for the advection term. However, a secondary objective of this research is to develop a means to estimate mesoscale vertical motion fields when one has soundings with mesoscale resolution in space and time, but with too few sounding sites to constitute a network adequate for computations with either of (4.1) or (4.4). Specifically, the Alberta Hail Project (AHP) is not equipped to operate more than half a dozen upper air sites simultaneously, but it can, and often does conduct sequential soundings from several sites on severe hailstorm days. In such cases, the lack of sampling points virtually rules out the use of the kinematic method for vertical velocity computations because of the requirement of an accurate and detailed mesoscale wind field.

The lack of spatial resolution also presents a problem in using the adiabatic method since temperature advection is required. One way around the problem might be to use synoptic scale data to estimate the advection term in (4.4). However, this may be grossly different from the mesoscale advection, especially since it would ordinarily be limited to 12-hour temporal resolution. Another possibility is to omit the horizontal advection term in (4.3) in the absence of fast-moving fronts. Also, if we apply the approximation

$$\omega = -\rho g w \quad (4.7)$$

to the RHS of (4.3), we obtain the result,

$$w \approx -\frac{1}{\Gamma_d} \frac{d_z T}{dt} \quad (4.8)$$

where

$$\frac{d_z T}{dt} = \frac{\partial T}{\partial t} + \omega \frac{\partial T}{\partial p} \quad (4.9)$$

Equation (4.8) is a quasi-Lagrangian approximation to (4.4), where $d_z T/dt$ is the rate of change of temperature following a parcel in the vertical, but neglecting its horizontal motion. The evaluation of (4.9) is depicted in Figure 4.4a, representing portions of two consecutive temperature soundings at the same site. The partial rate of temperature change, $\partial T/\partial t$, is simply the difference in sounding temperatures at pressure p_0 , divided by the time interval. The quasi-Lagrangian temperature change, (4.9), is the difference obtained by raising a parcel of air dry-adiabatically from temperature $T_0(A)$ at pressure p_0 , to $T_0(C)$ at pressure p_1 . Before applying (4.8), one must justify the approximation of negligible advection, or otherwise at least estimate its effect on the vertical velocity computation. Figure 4.4b shows the results of continued mid-level adiabatic ascent on a series of actual AVE-V soundings.

Since two sounding times must be used to compute a vertical velocity profile at a site, the time of the computed profile was assumed to be the average of the two sounding times in the cases to follow. The scheduled releases for SESAME-AVE were at 3-hour intervals, commencing at 1100 UTC, and continuing for 24 hours. Vertical velocity times are therefore 1230 UTC, 1530 UTC, and so on. In cases where the average vertical velocity time for a site exceeded these times by more than 1/2

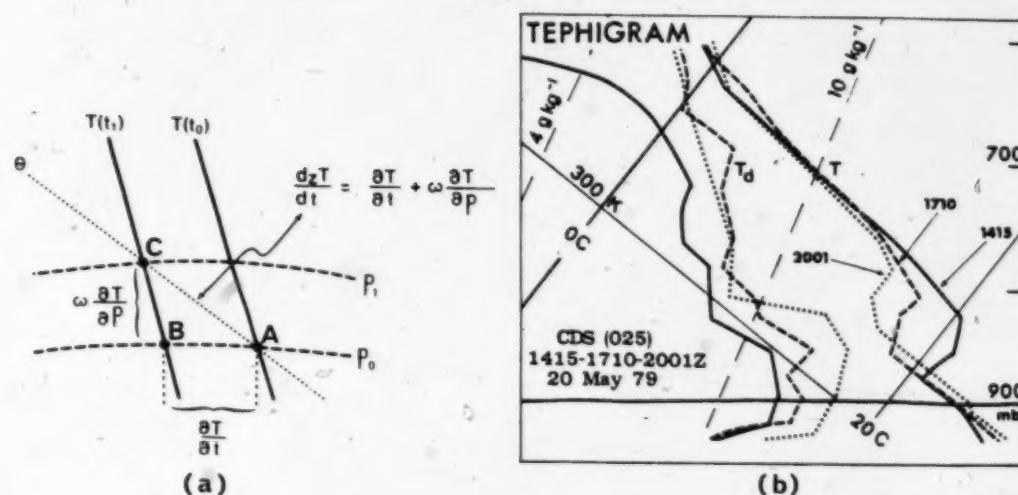


FIGURE 4.4: Sequential modification of soundings at a single site through adiabatic vertical motion; (a) schematic representation of modified adiabatic computations; (b) example for Childress, Texas, 20 May, 1979.

hour, computations were omitted from the final analyses in order to avoid temporal resolution problems. Additional soundings by mesoscale sites at the added 2130 UTC release time (convective storm maturity) were also omitted from the vertical velocity computations, in order to retain the same three-hour temporal resolution throughout the analysis.

4.4 INTERPRETATION AND VALIDATION OF VERTICAL VELOCITY COMPUTATIONS

Except for boundary layer turbulence work, atmospheric vertical motion is not at present a directly measured quantity. We have seen that not only must it be computed, but that the computational methods are somewhat imprecise. It is therefore important to consider how the

computed results should be interpreted, and what restrictions should be placed on these interpretations.

4.4.1 Interpretation of Kinematic Computations

In the kinematic method of (4.1), the vertical velocity is computed from the divergence of the wind field, which depends on (usually) small deviations from geostrophic wind conditions. This requires a detailed wind field, since, for centred-difference computations, four sounding sites are necessary for one value of divergence. This has the added effect of smoothing the vertical velocity field spatially, so that important features may be filtered out. More importantly, however, the wind field must be accurate, since relatively minor errors in the wind field can lead to significant errors in computed vertical velocity, as indicated earlier.

The latter problem raises the question of how convective storms might alter the wind field, and hence alter computations of vertical motion. It is known that convection does enhance the environmental vertical motion field locally once a storm forms. We need to consider how this might affect kinematic computations. Studies of severe storm environments have provided extreme kinematic vertical motion values of $20\text{--}80\text{ cm s}^{-1}$ (e.g., Fankhauser, 1969; Ogura and Chen, 1977; Fuelberg and Printy, 1983). It is not clear whether such values are truly representative of the environmental (as opposed to storm) field, although the rawinsondes providing these data are released in the environment, and not into storms. In fact, data from sondes which penetrate convective cells, almost always yield erratic data, and are

normally removed from data sets such as the SESAME data (Sienkiewicz et al., 1981). One should therefore be cautious when interpreting kinematic computations of severe storm environments. This problem has been addressed little in the literature, so that it requires some additional consideration at this point.

Satellite, doppler radar, and aircraft data show that the environmental wind field around storms is significantly altered (e.g., Lemon and Doswell, 1979; Houze, 1981). A storm acts as a partial block to the environmental flow, so that some of the air flows through and some of it flows around the storm. The blocking effect means that even winds at some distance from the cloud complex are affected. This does not invalidate kinematic computations of vertical motion. However, some of the interpretations in the literature may be incorrect, since the values are often referred to as environmental vertical motions.

Because only data from sondes which remain out of convective cells are retained, rawinsonde measurements are therefore supposed to be representative only for the extra-storm environmental conditions. In practice of course, the wind data around these storms is not purely environmental flow. In fact, many authors correctly state that their kinematic values include the storm scale motion, even though "pure" rawinsonde data is not supposed to sample convective storms.

On the other hand, thermodynamic variables should not be affected by this storm blocking, and should therefore be sampling the true environmental conditions, except when they penetrate the convective clouds. Supporting this, the adiabatic "environmental" vertical velocities in this study retain about the same range of values

throughout the storm complex life cycle, with absolute values generally $< 15 \text{ cm s}^{-1}$. On the other hand, the kinematic values for this case approach 100 cm s^{-1} during the mature storm period (Fuelberg and Printy, 1983). Assuming that both the thermodynamic and wind measures are correct, then the differences between the two methods must be reconciled, or else one or both methods must be declared invalid.

In fact, the solution to this problem is that both methods are basically correct, within the limits of data uncertainty. However, because the wind field is affected at great distances from the convective complex, when one uses the kinematic method, the combined environmental and storm-scale vertical motion field is effectively averaged over distances dependent on rawinsonde site spacing (80 km for the mesoscale, or 370 km for the synoptic network). This places two restrictions on its use: first, many mesoscale regions of subsidence within the blocking influence of a storm may not be resolved, because storm-scale ascent is averaged out over the larger region and swamps the vertical motion field in its immediate vicinity; second, point computations of vertical motion within this affected storm environment region will not be correct for the very same reason.

The adiabatic technique, used to compute vertical motion for the current study, is therefore believed to provide better estimates of the extra-storm (environmental only) component of the vertical motion field. This statement will be supported by direct comparisons towards the end of this chapter.

4.4.2 Interpretation of Adiabatic Computations

The adiabatic method of (4.4) has three main restrictions. First, it is valid only for unsaturated ascent, and it should therefore be applied with caution in the vicinity of clouds. However, this equation can be restated for saturated ascent by replacing the constant Γ_d with the (variable) saturated lapse rate, Γ_s (Petterssen, 1956; Hansen and Thompson, 1965). Second, uncertainties arise in the computation when the atmospheric lapse rate is close to adiabatic, because (4.6) then becomes relatively small. Third, the adiabatic technique does not account for any diabatic effects directly, which involve almost exclusively diurnal heating and cooling. In using the modified adiabatic technique of (4.8), a fourth restriction is that one must justify the neglect of advection.

Since diabatic processes are not included directly in the computations, their absolute effect can be lessened by decreasing the time interval between soundings. In order to decide on a critical sounding time interval beyond which surface heating becomes important, the July diurnal temperature trends (two-year average) for the surface (based on the difference between mean maxima/minima) and several mandatory levels (based on the difference between mean 0000 and 1200 UTC values) for Red Deer, Alberta were plotted in Figure 4.5a. Normally, the diurnal heating rate is a maximum during the morning for the boundary layer (generally below 800 mb), and during the afternoon at mid-levels (700 mb). However, only two soundings per day (1200 and 0000 UTC) were available for this estimation. Nevertheless, the approximation of linear diurnal temperature trends on the average (Figure 4.5a) are valid for this purpose.

The absolute uncertainty in a rawinsonde temperature is $\pm 0.5^\circ\text{C}$ (Sackiw and Strong, 1983). If this uncertainty is assumed to be the critical temperature change to be considered as significant diurnal heating or cooling, then the critical sounding interval is 12 hours for 500 mb, decreasing to about 3 hours for 750 mb, and only 1/2 hour for the surface (Figure 4.5b). The relation between critical sounding interval, Δt_c (in hours) and pressure level, is approximately linear below 500 mb according to

$$\Delta t_c = 0.5 - 0.0175(p - p_s) \quad (4.10)$$

where p_s is surface pressure, generally 1000-900 mb for most of the data sets used in this study. It is apparent that the adiabatic method will yield uncertain results for synoptic scale analyses below 500 mb, where the sounding time interval is 12 hours or greater, unless diabatic effects are included directly.

Figure 4.5c illustrates how surface heating changed the lower portion of an actual morning sounding exhibiting an inversion-capped boundary layer, prior to a severe Alberta hailstorm case (1330 UTC, 16 August, 1973). The surface heating trend was deduced from hourly surface temperatures at the sounding site. For cases such as this, there is virtually no diabatic heating above the stable layer before late afternoon. Thus, the diabatic heating rate averaged through much of the day can often be considered negligible above 800 mb. If uncorrected below 800 mb, however, vertical velocity computations will be underestimated for ascent, and overestimated for subsidence cases, all other factors being equal.

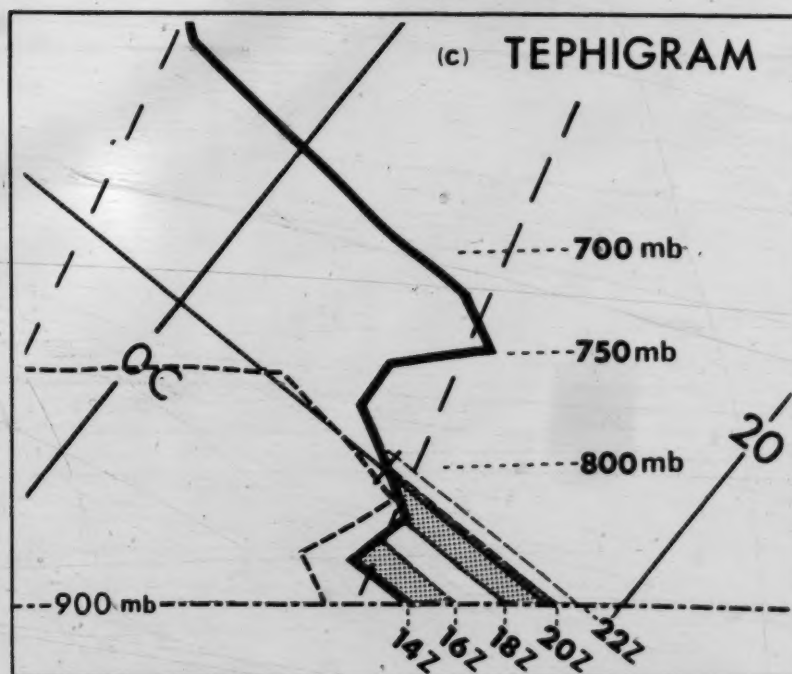
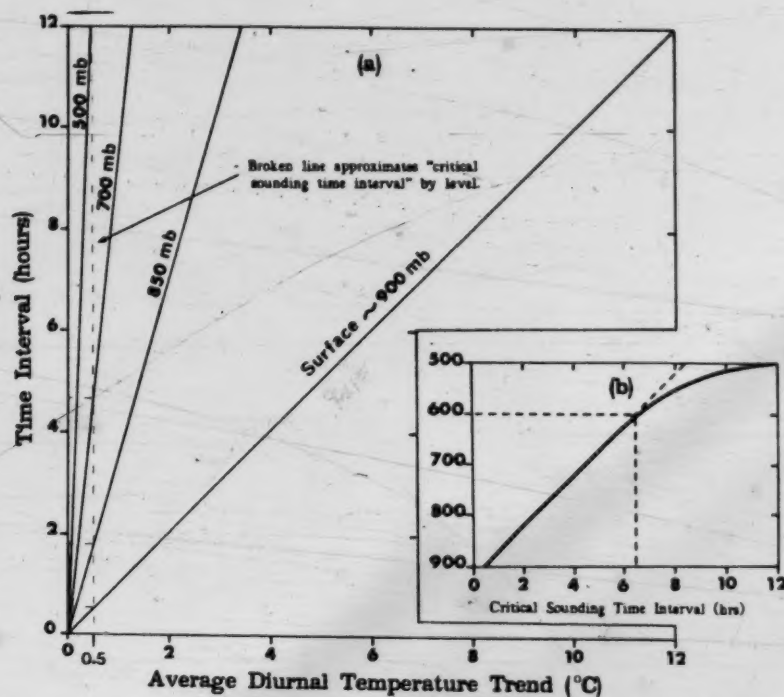


FIGURE 4.5: (a) Average (2-year) diurnal temperature trend by level during July at Red Deer Airport, Alberta; (b) approximate critical sounding time intervals by pressure level, required for neglect of diurnal heating using the adiabatic method; (c) actual surface heating trend on a morning sounding for Rocky Mountain House, 16 August, 1973, illustrating no diabatic heating trend above an inversion-capped boundary layer.

4.4.3 Application Restrictions

To summarize, the kinematic relation (Equation 4.1) is prone to errors in the wind field, but these can be filtered in most synoptic scale applications. When rawinsonde data from sites in the vicinity of thunderstorm complexes are included, however, the environmental winds can be, and often are, altered by these complexes. Then, by using the kinematic method, even the synoptic scale vertical motion field can become swamped by the thunderstorm flow field. However, because the rawinsonde site spacing, even for mesoscale networks such as SESAME, is still large compared with the scale of thunderstorm vertical motions, these 'storm scale' motions are effectively averaged over unrepresentatively large areas. While the composite effect on the synoptic scale may be unaffected, the computed vertical motion field in these cases do not reflect the true field either within or around the convective complex. This raises some doubt as to the validity of interpretation of kinematic vertical motions fields in such instances.

The adiabatic method of Equation (4.2) is less sensitive to wind errors, while (4.8) does not depend on the wind field at all. Furthermore, the environmental temperature field should not be affected by convective storms to the same extent as the wind field, since its response is not a direct dynamic one such as the wind deflection problem. In fact, Wiin-Nielsen (1964) has shown that energy conversion calculations, using adiabatic computations of vertical velocity, yielded only those components of energy conversion (between APE and KE) "not due" to the release of latent heat. Thus, for some applications the adiabatic method is preferable for mesoscale applications involving severe storms, where one is primarily interested in processes external

to the storm; in other words, with latent heat effects removed. Used for this purpose, one should expect adiabatic computations of ascent to be generally smaller than kinematic estimates in the vicinity of severe storms. Because of this, and so long as the sondes involved have not penetrated active convective cells, adiabatic computations should reflect more accurately the environmental vertical velocity field, whereas kinematic computations become dominated by storm scale motions.

Unless corrections can be applied for surface heating and cooling, however, the adiabatic method may be unreliable for boundary layer computations, or where sounding intervals are much greater than three hours. Also, the utmost care must be taken when this method is used in the vicinity of clouds, unless corrections are applied for saturated ascent. Without corrections, both diabatic heating and cloud penetration will lead to underestimates of ascent rates or overestimates of subsidence, but they will not usually affect the sign of the vertical motion. Nocturnal cooling has a negligible effect on vertical motion computations above 850 mb, since unlike diurnal heating, the cooling effect of surface radiation at night does not extend appreciably above the lowest 500 meters (~50 mb).

4.5 ERROR ANALYSIS AND FILTERING OF MODIFIED ADIABATIC COMPUTATIONS

Some of the difficulties associated with the computation of vertical motion have been reviewed. All methods used to compute this variable from rawinsonde data must use some or all of the pressure, thermodynamic, and wind data. The latter is itself a derived variable, obtained by tracking the rawinsonde balloon instrument, using either

radio, radar, or optical tracking methods. Together, these data comprise the atmospheric "sounding". The accuracy of sounding variables are now examined briefly, in order to arrive at estimates of the uncertainty in the modified adiabatic computations of vertical motion.

4.5.1 Sounding Accuracy

Estimates of the RMS errors, sometimes referred to as data uncertainty, in the SESAME rawinsonde data have been provided by Sienkiewicz et al. (1981), and are reproduced here in Table 4.1. Similar estimates have been reported by others, including Kurihara (1961), Danielsen and Duquet (1967), Fuelberg (1974), Richner and Phillips (1981), and Sackiw and Strong (1983).

TABLE 4.1: Estimates of RMS errors in SESAME rawinsonde data as reported by Sienkiewicz et al. (1981).

Temperature	0.5°C, all levels			
Pressure	1.3 mb, Surface to 400 mb 1.1 mb, 400-100 mb 0.7 mb, 100- 10 mb.			
Humidity	10%, not computed if $T < -40^{\circ}\text{C}$			
Pressure Altitude	10 gpm at 500 mb 20 gpm at 300 mb 50 gpm at 50 mb			
Winds				
	DIRECTION (degrees)		SPEED (m s^{-1})	
	at 10 degs.	40 degs.	10 degs.	40 degs.
Pressure	elevation	elevation	elevation	elevation
700 mb	9.5	1.3	2.5	0.5
500 mb	13.4	1.8	4.5	0.8
300 mb	18.0	2.5	7.8	1.0

These are realistic error estimates for normal operational conditions, but they are not appropriate for extreme cases caused by observer baseline errors, poor operational logistics, or systematic errors due to a poorly calibrated sonde. For example, a systematic temperature calibration error of only 1C, will lead to pressure altitude errors exceeding 50 gpm above 200 mb (Fuelberg, 1974). An incorrect baseline pressure of 1 mb can have a similar effect, while sondes which penetrate active convective cells sometimes yield quite erratic data.

The SESAME data sets underwent extensive quality control and error checking and most errors were either corrected or the sounding was flagged as having abnormal characteristics (Sienkiewicz et al., 1981). Sienkiewicz et al. (1981) provide a complete listing of the SESAME-AVE-V data at 25-mb intervals, along with an inventory of missing sounding data and those soundings with abnormal characteristics. The latter were omitted in all the analyses which follow in Chapters 5, 6, and 7.

4.5.2 Error Sources in the Modified Adiabatic Method

The main sources of error to be considered when computing the vertical velocity field using Equation 4.8 are:

1. the assumption of negligible temperature advection;
2. the assumption of no diabatic effects;
3. sensor inaccuracies and observer errors;

4. latent heat errors due to the rawinsonde passing through cloud layers;
5. differences in rawinsonde release times between sites;
6. horizontal drift of the rawinsonde balloon due to the winds.

These error sources are now examined for the major case study of this research, the SESAME-AVE-V case of 20-21 May, 1979 over Oklahoma and northern Texas. The first two sources, advection and diabatic heating, are examined separately because of their potentially large impact on vertical motion computations.

4.5.3 Data Uncertainties due to Advection and Diabatic Heating

Mean temperature changes due to advection and diabatic heating, and the average observed temperature changes at 850, 700, and 500 mb over the centre of the SESAME mesoscale network, are summarized in Table 4.2. The net change (last column of Table 4.2) represents temperature changes due to other processes. The advection and observed temperature changes are simply average rates, which were estimated from the the 1200 UTC synoptic analyses (Figures 5.1-5.2, to be discussed in Chapter 5). For comparison, a portion of the 0000 UTC 700 mb mesoscale analysis is shown in Figure 4.6. The 12-hour values were averaged to 3-hour changes in order to conform to the sounding time interval used in the adiabatic vertical velocity computations. The diabatic heating rates are estimates from the climatological statistics of Section 4.4.2 and Figure 4.5a for Alberta, since comparable statistics for Oklahoma were not readily available.

TABLE 4.2: Average 3-hour temperature changes (1200-0000 UTC) for central Oklahoma due to synoptic scale advection estimated at 1200 UTC, 20 May 1979, the average diabatic heating trend (Alberta climatology), the observed temperature change, and the net change due to other processes.

Pressure Level (mb)	Three-hour Temperature Changes			Net Change Attributed to Vert. Motion
	Projected Advection	Climatological Diabatic	Median Observed	
500	+ 0.29	+ 0.13	+ 0.40	- 0.02 °C
700	+ 0.43	+ 0.33	- 0.85	- 1.61 °C
850	+ 0.54	+ 0.88	- 1.10	- 2.52 °C

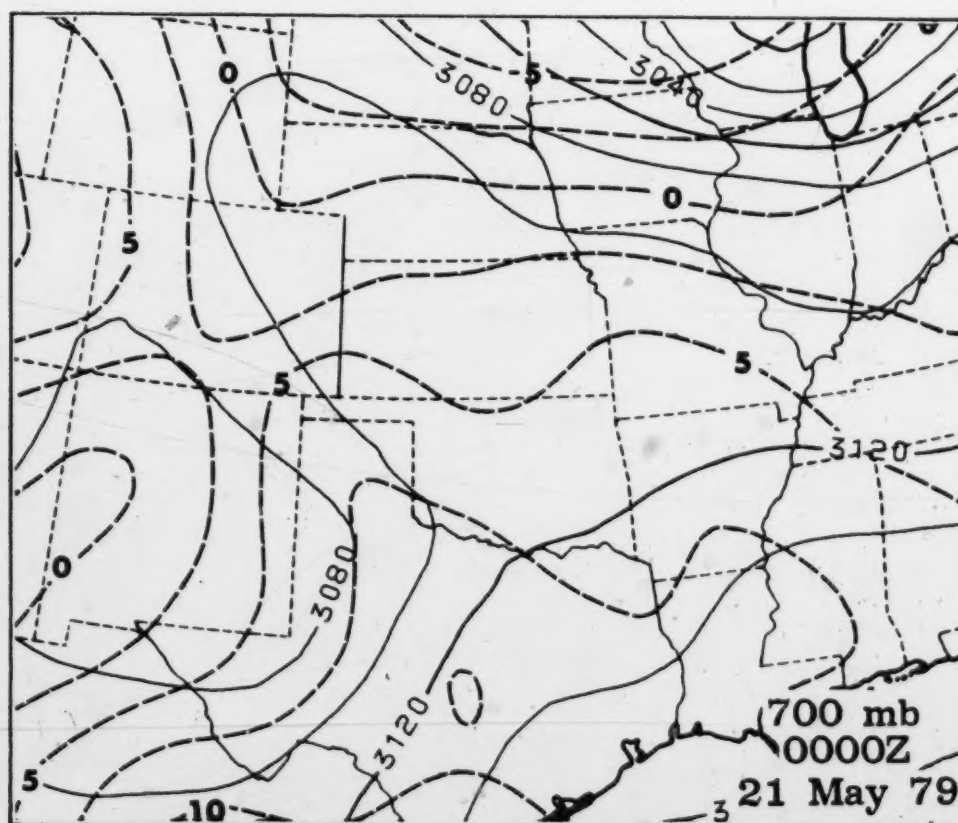


FIGURE 4.6: 700 mb analysis of heights (in m) and temperatures (in °C) for 0000 UTC, 21 May, 1979, using available synoptic and mesoscale data.

Using actual temperatures and winds, the synoptic analyses indicated advective warming at all three levels at 1200 UTC, with a light (850 mb) to moderate (500 mb) southwest flow aloft. This amounted to about $+2^{\circ}\text{C}$ at 850 and 700mb, and the trend persisted at 0000 UTC. The advective warming must be added to the normal (climatological) diabatic heating from insolation or other surface effects. However, in spite of these additive effects of warming, significant cooling occurred at both 850 and 700 mb. Significant adiabatic ascent appears to be the explanation for the net changes (last column of Table 4.2). In other words, the adiabatic vertical velocity values were based on the observed temperature changes of Table 4.2, but if advection and diabatic heating had been incorporated, the computations would have used values comparable to the last column of the table.

In practice, it is difficult to incorporate all of these effects in computing adiabatic values of vertical motion, especially that of diabatic heating. The greatest proportion of the latter effect results from surface heating due to insolation, and this is not distributed evenly throughout the troposphere, especially when one must contend with low-level temperature inversions as in this case. The heating trend provided with the capping lid sounding of Figure 4.5c shows that the 850 mb level would heat more rapidly during the morning than Table 4.2 indicates, while the 700 and 500 mb levels probably would be unaffected initially. Once the inversion disappears, surface heat is then distributed more uniformly to these upper levels. Thus, the average values of Table 4.2 must be considered with caution. At the same time, the small rates of temperature change in Table 4.2 (changes of 0.5°C or less) are difficult to resolve at the mesoscale within the 3-hour

sounding release schedule, since these values are of the order of, or less than the RMS error (0.5°C) for rawinsonde data (Table 4.1). Nevertheless, actual net three-hour cooling rates of $1\text{--}3^{\circ}\text{C}$ were common for this SESAME case, especially in the 850-700 mb layer where the capping inversion was a common feature (see the sequential soundings depicted in Figures 5.23, 6.14, 6.15, 6.23, and 6.24).

Weak thermal advection is evident in Figure 4.6, comparable to that of the synoptic scale analysis, except over the northern Texas Panhandle and at the southern Oklahoma border near the rawinsonde site at Wichita Falls (SPS). The manually-estimated advection rate for the latter region was $+1^{\circ}\text{C}/3\text{ hr}$ at 850 mb, certainly of resolvable magnitude. This value was the maximum advective rate of warming for this case, and was counteracting cooling due to adiabatic ascent (last column in Table 4.2) by a maximum of 40%. The resulting maximum error in computed adiabatic ascent is less than 4 cm s^{-1} .

The observed 12-hour (1200 to 2400 UTC) temperature changes for 850, 700 and 500 mb, using both synoptic and mesoscale data, are shown in Figure 4.7. Clearly, vertical motion (ascent) was more than compensating both advective warming and diabatic heating from insolation throughout the period below 500 mb. A line of maximum cooling across southern Oklahoma at 850 mb appears to be displaced further north with height (Figure 4.7). This cooling pattern was mainly a result of adiabatic ascent, and may have been a factor in the southward propagation of the surface cold front, despite the southerly wind component on its south side (see Figures 5.1-5.2). While the foregoing demonstrates that the advection term is not unimportant, it appears that the environmental component of vertical motion was the main influence on

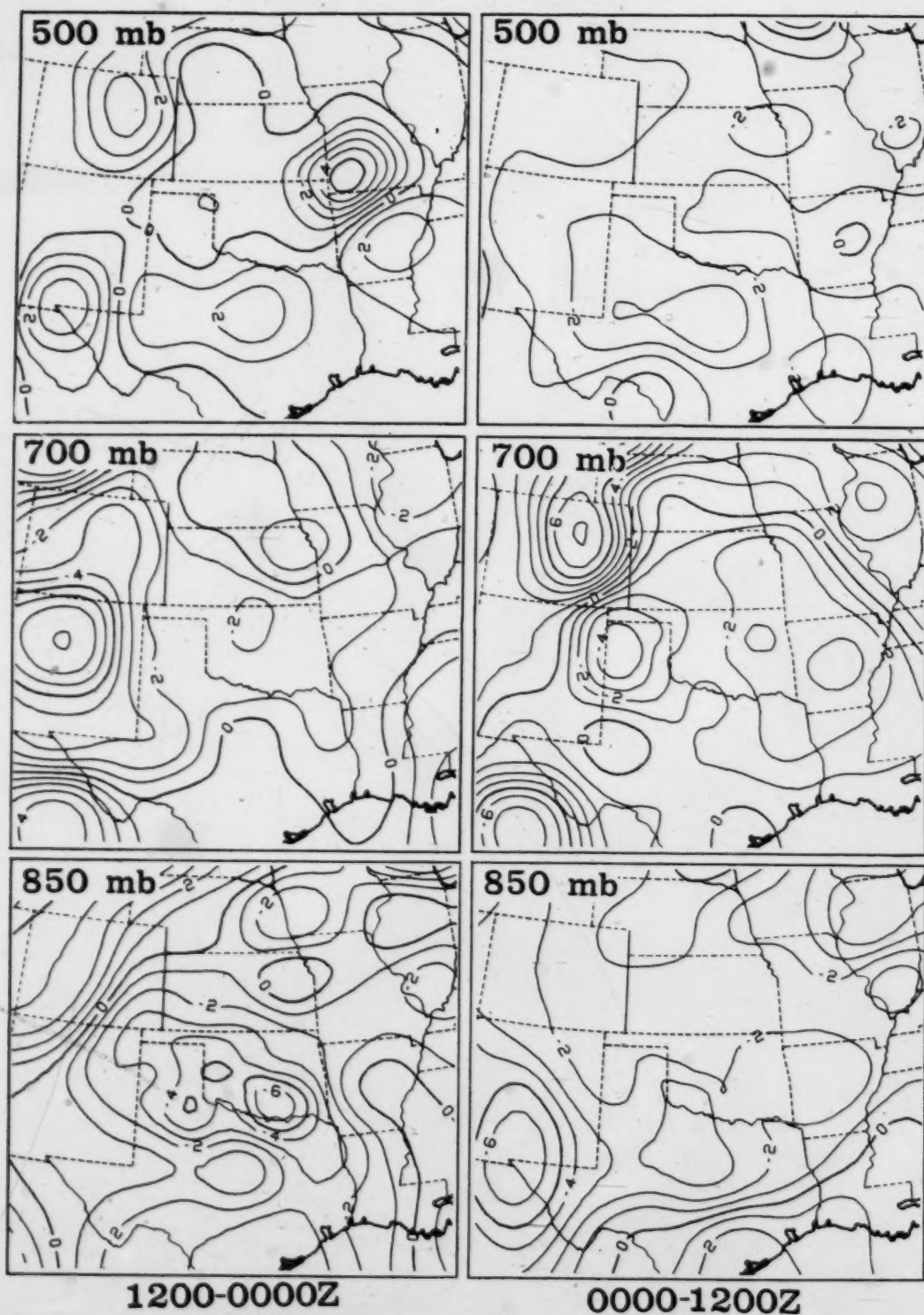


FIGURE 4.7: Net temperature changes ($^{\circ}\text{C}$) at 500, 700, and 850 mb during the 12-hour periods, 1200-0000 and 0000-1200 UTC, 20-21 May, 1979.

the observed sounding temperature changes in this case.

4.5.4 Error Analysis Summary, Known Sources

The uncertainty in the computed values of vertical velocity due to the six sources listed in Section 4.4.2 can now be summarized. Using a range of advection values (for the AVE-V case) of $+0.3$ to $+1.0$ $^{\circ}\text{C}/3$ hr, its exclusion from Equation 4.8 leads to an uncertainty in the vertical velocity of -0.3 to 4.0 cm s^{-1} . Failure to account for diabatic surface heating could cause an underestimate of cooling due to ascent of up to $4^{\circ}\text{C}/12$ hr below 850 mb, and a further error in 'w' of up to 1.5 cm s^{-1} . At 700 mb, the maximum diabatic (heating) error contribution to 'w' is <0.4 cm s^{-1} . The uncertainty in 'w' due to sensor inaccuracies is ± 0.5 cm s^{-1} , based on the temperature uncertainty of $\pm 0.5^{\circ}\text{C}$ (Table 4.1).

Errors due to rawinsonde cloud penetration are assumed to be <0.5 cm s^{-1} , since those portions of the vertical velocity profiles where saturation occurred were removed from the data set. Saturation could occur either from a brief balloon penetration of clouds, or because theoretical saturation was achieved as the lower portion of the sounding was lifted dry adiabatically. The latter could have been handled better by assuming the pseudo-adiabatic cooling rate from the point where theoretical saturation was achieved, instead of removing the data.

Occasionally, operational logistics force balloon releases off schedule. As pointed out in Section 4.2.4, the time of a vertical velocity profile was assumed to be the average release time between the

two consecutive soundings used. If this was more than one half hour off the average scheduled times, then that profile was removed from the final results. The uncertainty in the vertical velocity due to the average release time being out by one-half hour, is $< 1.0 \text{ cm s}^{-1}$.

The highest pressure level used in the isobaric analyses to follow was 600 mb, primarily because this study is concerned mainly with the pre-storm sub-cloud layer, below 600 mb. The ascent time to this level, using the accepted standard balloon ascent rate of 5 m s^{-1} , is approximately 14 minutes. Slight variations of a minute or so from this ascent time, would not significantly affect the temporal or spatial resolution of the analyses. At the same time, balloon drift at all sites was less than 10 km at 600 mb (see sounding data provided by Sienkiewicz et al, 1981), causing no adverse displacement of vertical motion features up to this level, even for the mesoscale network with its 80-km spacing of sites. In any event, any displacement would be generally in the same direction, at least for neighbouring sites, so that at worst, the isobaric patterns (of Figures 5.6-5.16, for example) would be spatially shifted a few kilometers. Thus, the uncertainty in vertical velocity due to balloon drift is less than 1.0 cm s^{-1} , and far less than this below 700 mb.

The uncertainties in vertical velocity due to the above six error sources are summarized in Table 4.3. Vertical and temporal smoothing was carried out on the vertical velocity profiles. It will be shown that the spatial patterns in the analyses to follow are basically correct, while the absolute errors should not attain the maximum of almost 5 cm s^{-1} as shown in Table 4.3.

TABLE 4.3: Uncertainties in the modified adiabatic computations of vertical velocity for the SESAME-AVE-V case.

ERROR SOURCE	UNCERTAINTY (cm s^{-1})	
	Minimum	Maximum
1. Temperature Advection	- 0.3	+ 4.0
2. Diabatic Heating	0.0	+ 1.5
3. Sensor Inaccuracies	- 0.5	+ 0.5
4. Cloud Penetration	0.0	+ 0.5
5. Off-schedule Release Times	- 1.0	+ 1.0
6. Balloon Drift	- 1.0	+ 1.0
NET UNCERTAINTY*	± 1.5	± 4.6

* By definition (Topping, 1965), the net uncertainty is the square root of the sum of the squares.

4.5.5 Smoothing the Vertical Velocity Fields

Vertical velocity computations were carried out at 25 mb intervals for each sounding, terminating at 100 mb. The vertical velocity profiles were filtered in height, using a three-point linearly-weighted height filter with weights of 1, 2, and 1. Then, each set of vertically-smoothed profiles for each rawinsonde site was smoothed in time, using the same weighted filter at each level, to produce the final profiles. The filters were necessary to remove noise, but while reducing the amplitudes of the vertical velocity, they did not account for the differences in absolute values between the kinematic and adiabatic computations to be discussed in Section 4.6. It should be noted, however, that for the time filtering, it was necessary to use a

two-point filter at the end times, 201230 and 210930, with double weighting at these times. This had the effect of leaving slightly higher values at these times, compared with adjacent times, and tighter gradients in the isobaric contours.

4.6 COMPARISON OF KINEMATIC AND ADIABATIC COMPUTATIONS IN THIS STUDY

Fuelberg and Printy (1983, 1984) have computed detailed mesoscale and synoptic scale kinetic energy budgets for the AVE-V case. Fuelberg (personal communication) has supplied the author with some of his unpublished mesoscale kinematic analyses of vertical velocity at 700 mb. These will be compared with the results of the adiabatic method used here. The kinematic analyses are valid at the actual sounding times; e.g., 1100 and 1400 UTC. In order to make direct comparisons in time with Fuelberg's results, and to have temporal resolution comparable to that of the adiabatic analyses, Fuelberg's values were averaged over two adjacent times. Also, values of ' ω ' were converted to ' w ' using (4.7), with $\rho = 0.872 \times 10^{-3} \text{ g cm}^{-3}$ (for $P = 700 \text{ mb}$ and $T = 6.5^\circ\text{C}$) and $g = 979.7 \text{ m s}^{-2}$ (for 35°N latitude); i.e., $w = -1.17 \omega \text{ cm s}^{-1}$ for ω in $\mu\text{b s}^{-1}$.

Two comparisons, one for the pre-storm environment at 1230 UTC, the other for the mature convective storm stage at 2130 UTC, are shown in Figure 4.8. Fuelberg's kinematic analyses, hereafter referred to as K12 (1230 UTC) and K21 (2130 UTC), are based only on the mesoscale network data. The corresponding adiabatic analyses, referred to as A12 and A21, include all available data from the surrounding synoptic upper air stations as well. Therefore, since the scale of the disturbances

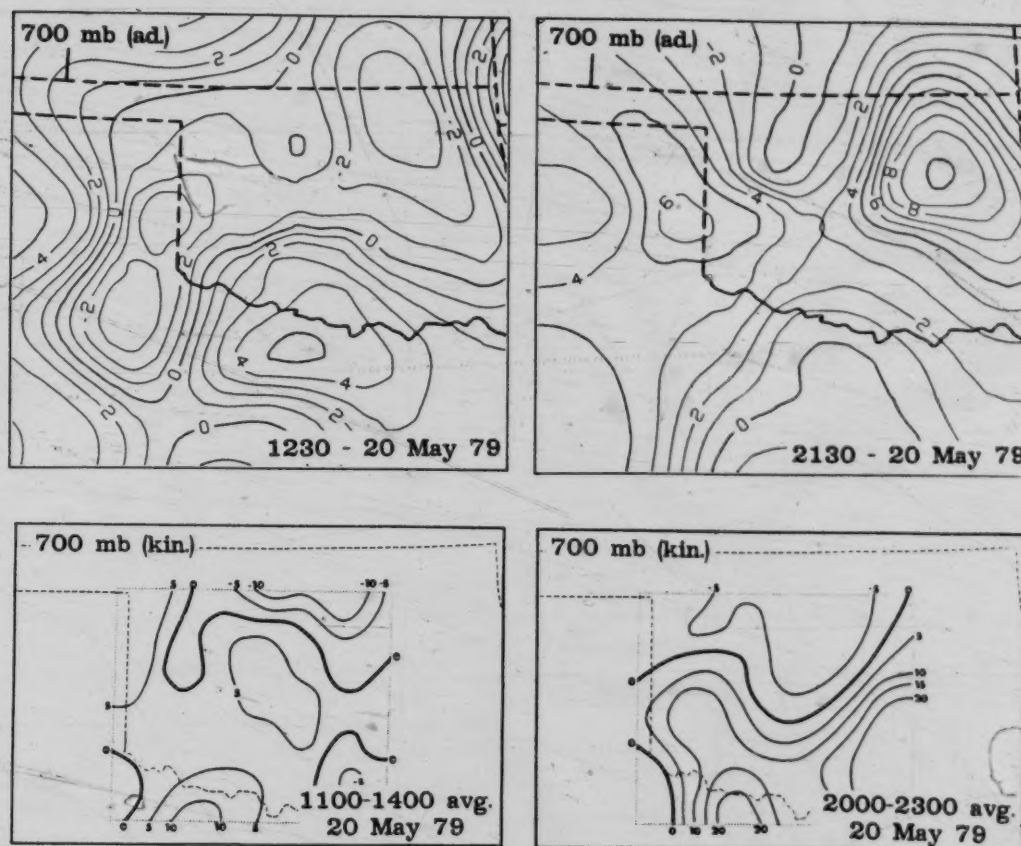


FIGURE 4.8: Comparative analyses of 700 mb vertical velocity (cm s^{-1}) for 1230 and 2130 UTC, 20 May, 1979, using (a) the modified adiabatic technique (ad.), and (b) the average of two kinematic analyses (kin.) by Fuelberg and Printy (1983 and personal communication). (Slightly different map projections and scales are used.)

(convective complexes) in the AVE-V case was slightly smaller than the size of the mesoscale network, one would expect the main difference in patterns to be around the edges of the network.

For the 1230 UTC comparison, the A12 maxima over southern Oklahoma and the Texas Panhandle west of the mesoscale network, and minima along the western Oklahoma border and over the northeast part of the state,

agree with similar centres on K12. The K12 centres, however, are displaced slightly from those of A12, perhaps because of edge effects in the objective analysis. However, a curious secondary maximum over the centre of the network on K12 has no such counterpart on A12; K12 values exceed 7 cm s^{-1} where A12 has negative values over this region. A qualitative comparison of A12 and K12 with the satellite imagery and radar composite of Figure 5.6 is most favourable toward the adiabatic analysis. It appears that the adiabatic computations lack time resolution, while the kinematic method suffers in spatial resolution. Both problems are functions of the types of computations involved.

By 2130 UTC, the GOES cloud-top imagery shows a major convective complex covering most of Oklahoma (Figure 5.11), while the radar composite shows that the southern half of Oklahoma is the most convectively-active. Major storm cells are evident near the southeast portion of the mesoscale network (radar tops of 16.7 km), the southwest portion (tops of 16.4 km), and both west (tops 14.2 km) and southwest (10.6 km) of the network. The mesoscale data resolution does not allow one to use either method to resolve vertical velocity patterns associated with all four of the above smaller scale major storm cells. Allowing for objective analysis edge effects (because the kinematic analysis includes only the mesoscale data), the sub-synoptic scale patterns of K21 and A21 are in good agreement with each other. On mesoscale detail, however, the A21 regions of ascent appear to describe mesoscale regions of cloud somewhat better. Also, in a qualitative comparison of cloud-free areas, as depicted by the satellite imagery, with the small regions of subsidence on A21 and the comparable K21 analysis, the adiabatic patterns compare better than the kinematic.

In spite of the fact that the two approaches to calculating 'w' are entirely different, and use a different data base (one dynamic, the other thermodynamic), the broad features in the patterns are very similar, although the ranges of values are different. Over the mesoscale network only, the A12 values range from -5 to +5 cm s^{-1} , while those of K12 are from -10 to +9 cm s^{-1} . For A21 the range is -1 to +10 cm s^{-1} , while the K21 range is -10 to +25 cm s^{-1} . The difference in the range of the values at 1230 UTC is -5 to +4 cm s^{-1} , close to the maximum net uncertainty in Table 4.3. However, at 2130 UTC, the difference between the kinematic and adiabatic approaches increases to -9 to +15 cm s^{-1} . These differences cannot be explained simply on the basis of the errors discussed in Table 4.3.

Fuelberg and Printy's (1983) values of vertical velocity for this case show a general increase with height, exceeding 60 cm s^{-1} at 400 mb during the period of storm maturity. But their values are in line with other studies which have used the kinematic method in the vicinity of severe storms. For example, Fankhauser (1969), Ogura and Chen (1977), and Park and Sikdar (1982), show vertical velocity values in the range of 10-60 cm s^{-1} near 700 mb. Pre-storm computations, however, as with Fuelberg's values, always tend to be under 10 cm s^{-1} . With the kinematic method, the computed vertical motion is dominated by the storm scale field. Kinematic computations, in particular, appear to be strongly influenced by the effect of the storm blocking and deflecting the wind field for considerable distances around the storm complex.

The adiabatic technique tends to retain values of vertical motion in the range of -10 to $+10 \text{ cm s}^{-1}$, both prior to and during the severe storm, although spot values of up to 20 cm s^{-1} do occur in the AVE-V case. These results therefore support the contention that the environmental thermodynamics are not as influenced by convective dynamics as the wind field is. Therefore, the adiabatic method may be better suited to estimate components of the vertical velocity field "external" to the convective processes. Because the kinematic method computes the "total" vertical velocity, including that due to storm-scale motion, the vertical velocity field around convective storms is smeared out (in effect, averaged) over too large an area, causing aliasing, and leaving results open to mis-interpretation.

4.7 A TEST OF FORECASTING CAPABILITY

The SESAME data sets provide a means to test a statistical convective forecasting routine, the Synoptic Index of Convection (Sc4), with extensive mesoscale data. The formulation of the Sc4 and its forerunner, the Sc10, are detailed in Appendix III for ease of reference, along with a summary of procedures for its use in forecasting convective complexes.

Simply put, the Sc4 is computed from two synoptic scale and two sub-synoptic scale input variables, each independently correlated to a predictand called the Convective Day-Category (CDC), defined in Table 4.4. The CDC was designed to be a single number which would represent the maximum degree of convective intensity occurring on a specified day over the ARC Hail Project operations area (AHP) of approximately 130 km

radius. Thus, it is important to realize that a point value of CDC (likewise, of the predictor, Sc4) represents conditions for a 130-km radius from that point, and actually represents the maximum convective intensity for the whole day, from 1700 to 1200 UTC the following morning, or 1100 to 0600 MDT.

TABLE 4.4: Definitions of Convective Day Categories for the AHP operations area (130-km radius from Red Deer Airport, Alberta).

CDC	DEFINITION
+ 5	LARGER THAN GOLFBALL, >52 mm diameter HAIL reported.
+ 4	GOLFBALL, 33-52 mm HAIL reported.
+ 3	WALNUT, 21-32 mm HAIL reported.
+ 2	GRAPE, 13-20 mm HAIL reported.
+ 1	PEA, 5-12 mm, or SHOT, <5 mm HAIL reported.
0	WIDESPREAD SHOWERS throughout most of the AHP area or the occurrence of a THUNDERSHOWER, but NO HAIL.
- 1	SCATTERED SHOWERS (over less than half the area), NO HAIL or THUNDERSHOWER.
- 2	SIGNIFICANT CONVECTION (ACC, TCU, or CB clouds) but NO SHOWERS.
- 3	NO DEEP CONVECTION reported (no ACC, TCU, or CB clouds).

NOTES: At least two hail reports are required to qualify for CDC > 0. All categories are determined from the period 1700 through 1200 UTC. See Strong (1979) or Strong and Wilson (1983a) for other qualifying remarks on these categories.

4.8 DISCUSSION

Any study of the interactions between synoptic processes and mesoscale convective storms, requires concurrent data at both scales of motion. The SESAME data sets are the most useful to date in meeting this need. The Montana CCOPE upper air data are useful for some mesoscale analysis, but they are totally unsuitable for scale interaction studies, while Alberta data have been limited to even smaller experiments with a very definite focus as demonstrated in Chapter 3. Therefore, qualitative comparisons were used to confirm that the severe convective storms of Alberta, Colorado, and Oklahoma involve common microphysical processes, and that they occur under similar circumstances of planetary and synoptic scale patterns. This was established in order to justify the later transfer of results from the SESAME mesoscale analyses back to Alberta hailstorm applications.

A simplified version of the adiabatic method for computing vertical motion was described. Its advantage is that it requires only two sequential soundings from a single upper air site in order to compute one vertical profile. The kinematic method, on the other hand, requires a spatial field of wind measurements (3-4 sites, depending on the computation technique) to produce one profile, since this technique is based on computations of the divergence field. Apart from this problem, wind measures are also quite error-prone. Thus, the adiabatic method was chosen because the technique could later be easily applied to Alberta storm cases, where only limited sounding sites are available.

Six possible error sources were identified in the adiabatic technique. Under the worst circumstances, the combined uncertainty in the computed vertical velocities could result in individual overestimates of subsidence of 1.5 cm s^{-1} , or underestimates of ascent of 5 cm s^{-1} . The actual errors are believed to be less.

More importantly, the adiabatic computations are believed to yield more representative values of the "environmental" component of vertical motion than do the kinematic computations. The main problem with the kinematic fields of vertical motion around severe storms appears to be in the interpretation of the results. Wind fields are known to be blocked and deflected over large regions in the vicinity of thunderstorms; that is, the winds are affected over regions perhaps several times larger than the storm itself. When these winds are used to compute the "environmental" portion of vertical motion, the result is really storm scale motion smeared out (averaged) over a large mesoscale region. In effect, aliasing occurs in the vertical motion pattern over the affected storm environment region. Since radiosonde thermodynamic data is representative only of the environmental conditions alone, the adiabatic method may therefore be better suited to estimate components of the vertical velocity field "external" to the convective processes. A qualitative comparison of the two methods, using the SESAME satellite imagery and radar composites, favours the adiabatic analyses over kinematic computations. Better methods of measuring or computing vertical motion, perhaps using Doppler radar, aircraft data, and techniques such as those described by Fankhauser et al. (1985), should be one goal of future mesoscale networks.

"Diagnostic studies are an essential means of forging the link of mutual understanding and respect between the theoretician generating the fundamentals of the science based on idealized models, and the weather analyst or forecaster attempting to apply his theories."

R.P. Pearce (1974)

CHAPTER 5

SESAME-AVE-V: 20-21 May, 1979 - A Case Study

The AVE-V study period was the fifth of six special SESAME case study days. During the first three cases, 20 additional rawinsonde systems were added to the existing synoptic network; for the last three, these 20 systems were combined into one mesoscale network as shown in Figure 4.2. Of the six SESAME data sets, the AVE-V period is the most complete in that a major convective complex passed right over the mesoscale network.

This chapter will focus on the pre-storm environment of the AVE-V period. Using synoptic and mesoscale analyses, an attempt is made to isolate synoptic scale dynamic factors which contributed to the formation of a severe convective storm over extreme southwest Oklahoma around 1830 UTC. The AVE-V case has been documented by only three other studies published to date. Fuelberg and Printy (1983) investigated the mesoscale kinetic energy balance within the confines of the special mesoscale network, while Printy and Fuelberg (1984) included a time- and area-averaged analysis of the synoptic scale kinetic energy balance. Both studies concentrated on the mature storm period, with some discussion of feedback mechanisms from the storm to the synoptic scale,

but they did not discuss possible downscale transfers (synoptic effects on the mesoscale), especially that during the pre-storm period. Sikdar and Fox (1983) computed mesoscale momentum budgets for this AVE-V case, and found a well-organized vertical circulation over the region at least 90 minutes prior to the earliest radar echoes, and low-level convergence four hours before the echoes.

In this chapter we investigate the vertical motion field in three dimensions, in order to detect relations between convective storm complexes and the pre-storm dynamics, as far in advance of storms as the SESAME data allows. The origins of the pre-storm ascent are discussed, including possible sources well upstream from the storms. The mechanisms by which such ascent might help to initiate the storms are reviewed in the context of the AVE-V case study.

5.1 SYNOPTIC SCALE SUMMARY

Because of the frequent references to analysis times in this and subsequent chapters, a convention is adopted here whereby analysis times will be indicated as "ddtttt", where "dd" is the day of the month and "tttt" the time in hours and minutes (UTC). The synoptic surface and upper air analyses for the AVE-V case (Figures 5.1-5.2) were obtained from CMC micro-film, and span the pre-storm period, 200000 - 201200, and the severe storm period, 210000 - 211200. The 250 mb jet streams and 850 mb wind streaks¹¹ are indicated by heavy arrows. These were added using objective analyses of isotachs at mandatory levels, taking

¹¹ Wind or jet streaks are simply lines of wind maxima (Uccellini and Johnson, 1979).

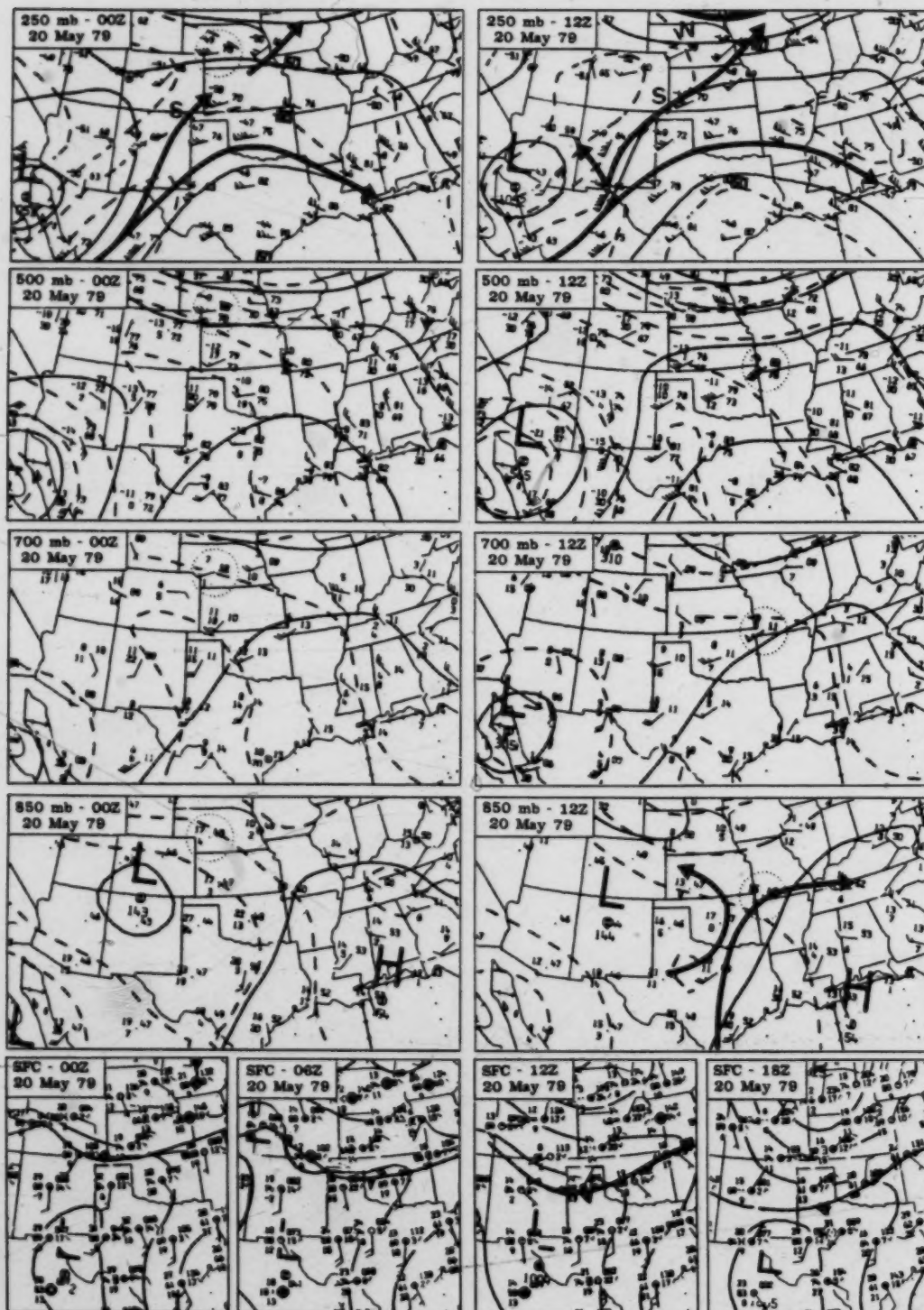


FIGURE 5.1: Synoptic Analyses: surface, 850, 700, 500, and 250 mb at 0000 and 1200 UTC, 20 May, 1979.

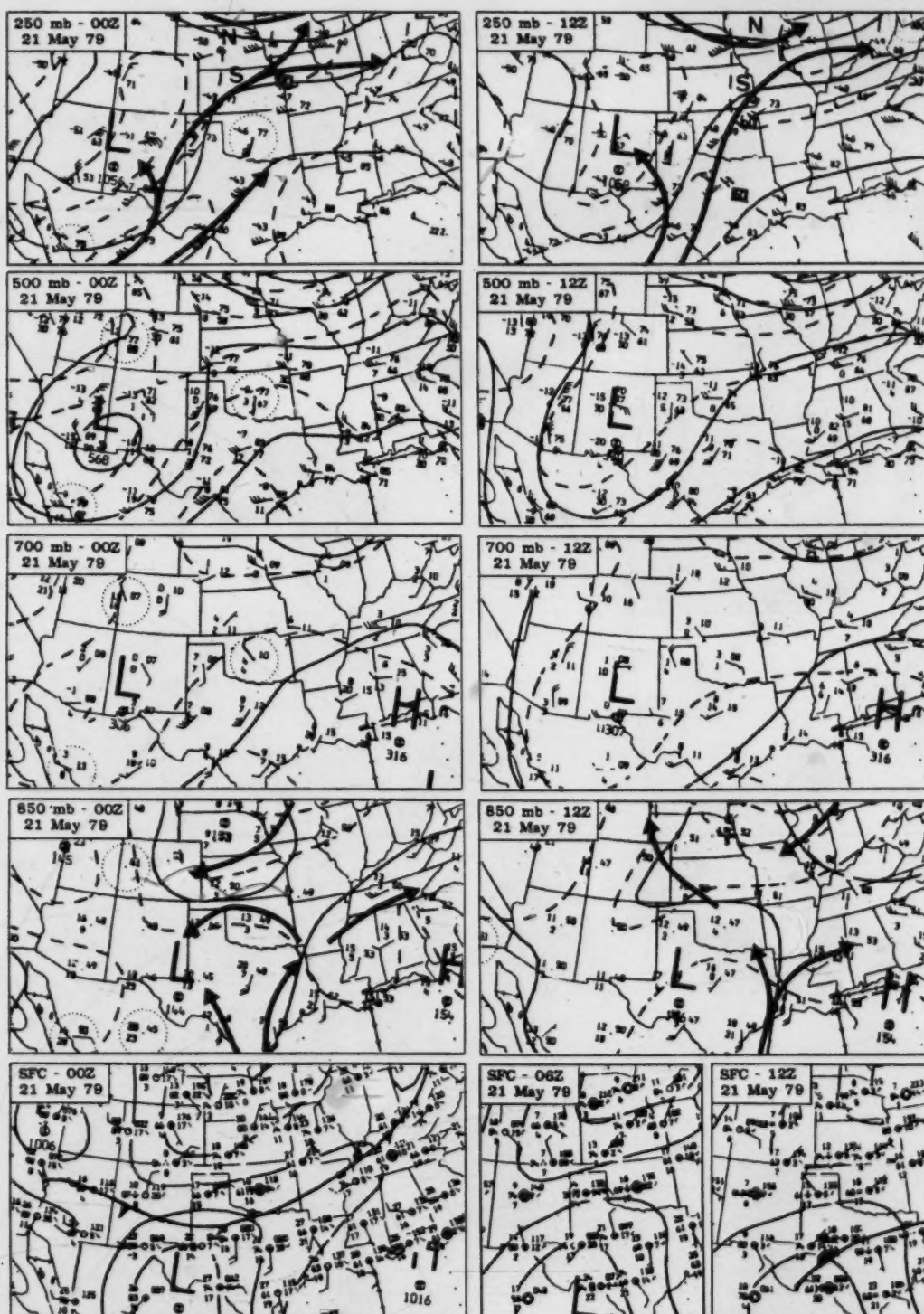


FIGURE 5.2: Synoptic Analyses: surface, 850, 700, 500, and 250 mb at 0000 and 1200 UTC, 21 May, 1979.

advantage of the additional mesoscale data of Appendix IVb. More detailed hourly analyses of the surface charts, using the additional mesoscale surface and upper air data, are presented in Appendix IVa.

The GOES visible images of Figure 5.3 give a synoptic scale perspective of the cloud systems and convective cloud complexes during the pre-storm (201301) and mature storm (202301) periods of the primary (AVE-V) convective complex over Oklahoma. Figure 5.4 provides schematic weather composite of synoptic scale cloud patterns, dynamic features, and convective complexes, at the two main synoptic hours, 1200 and 2400 UTC, and as the Oklahoma storm was forming around 1830 UTC. More detailed radar and satellite analyses are presented in the mesoscale analyses which follow in Section 5.2.

5.1.1 Synoptic Scale Kinematics

Throughout this storm period, eastern North America is dominated by two main cyclonic vortices, connected by an intermediate (planetary to synoptic) scale trough through the central U.S. These features are evident at all levels (Figures 5.1-5.2). One vortex is a quasi-stationary low centred over Hudson Bay (not shown), which results in a west to northwest flow across Canada and the northern U.S. The second vortex, a low over the Baja Peninsula at 200000, maintains a southwest flow over the southeast U.S. as it moves northeast into New Mexico during the decaying stage (after 210000) of the convective complexes.

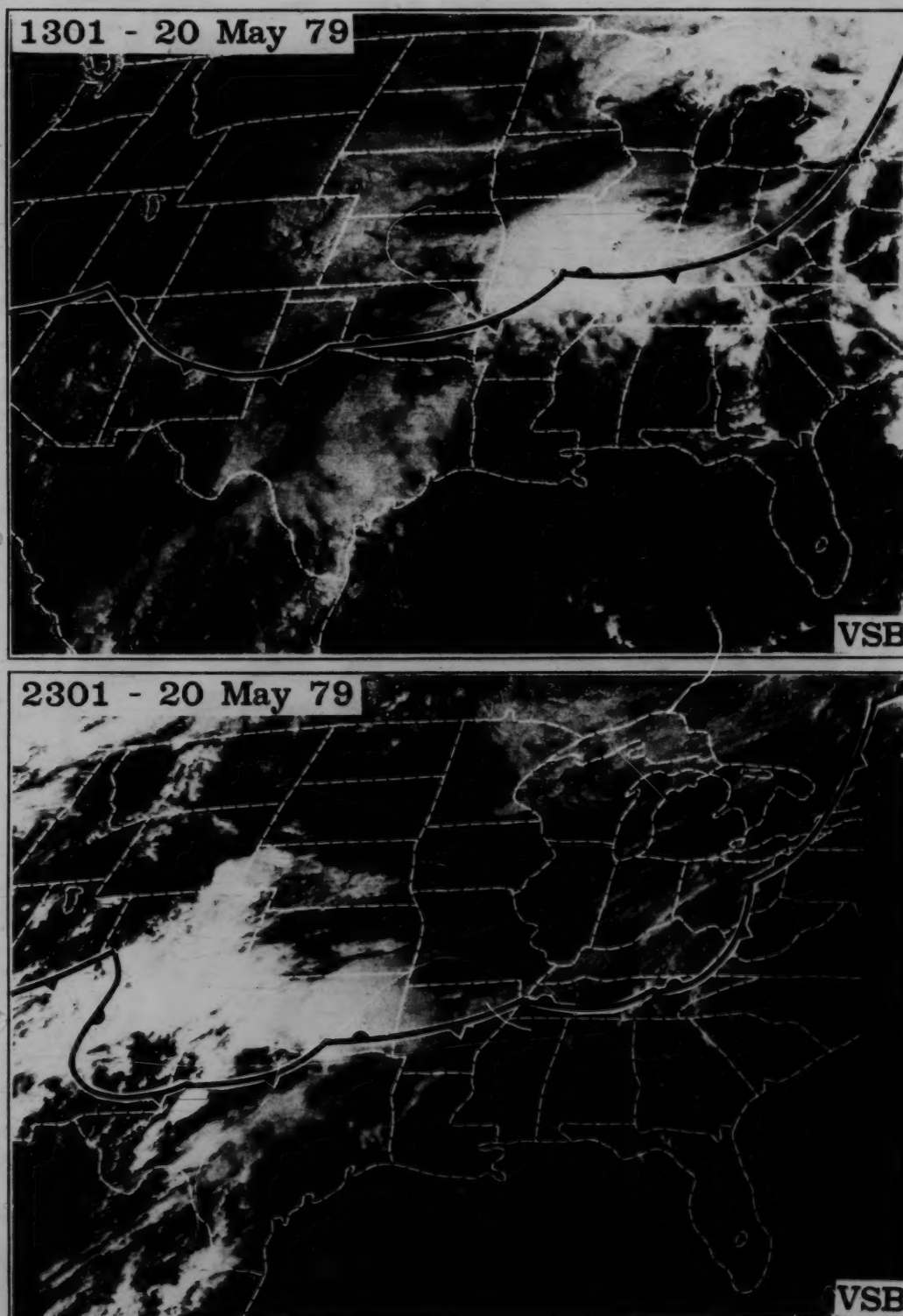


FIGURE 5.3: GOES-East visible imagery of cloud systems (synoptic perspective) during AVE-V, 20 May, 1979, for the (a) pre-storm period, 1301 UTC; (b) mature convective storm period, 2301 UTC. Superimposed fronts are from hourly surface analyses.

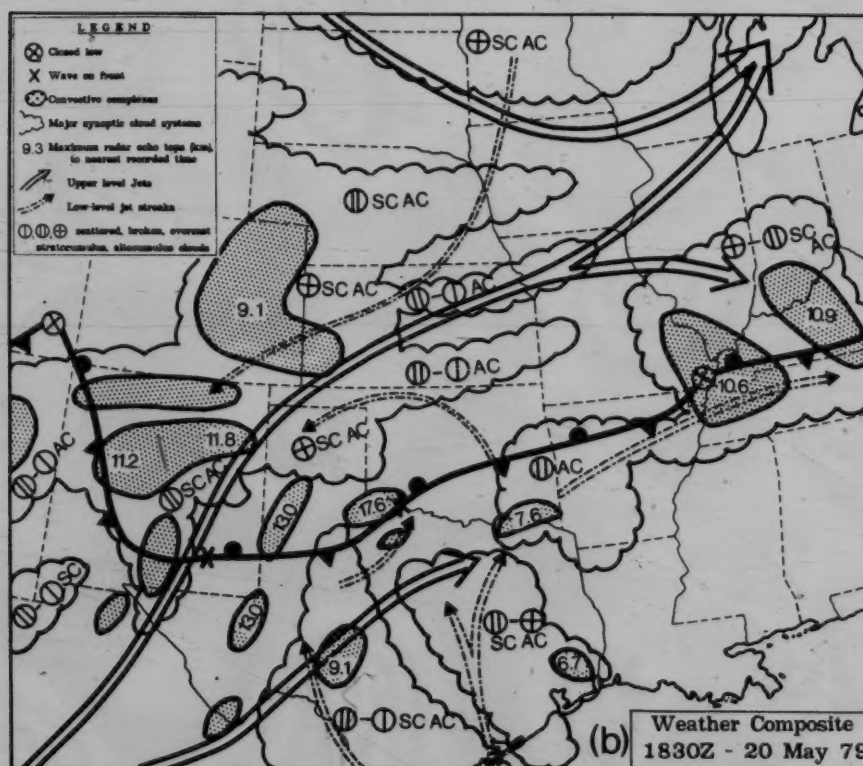
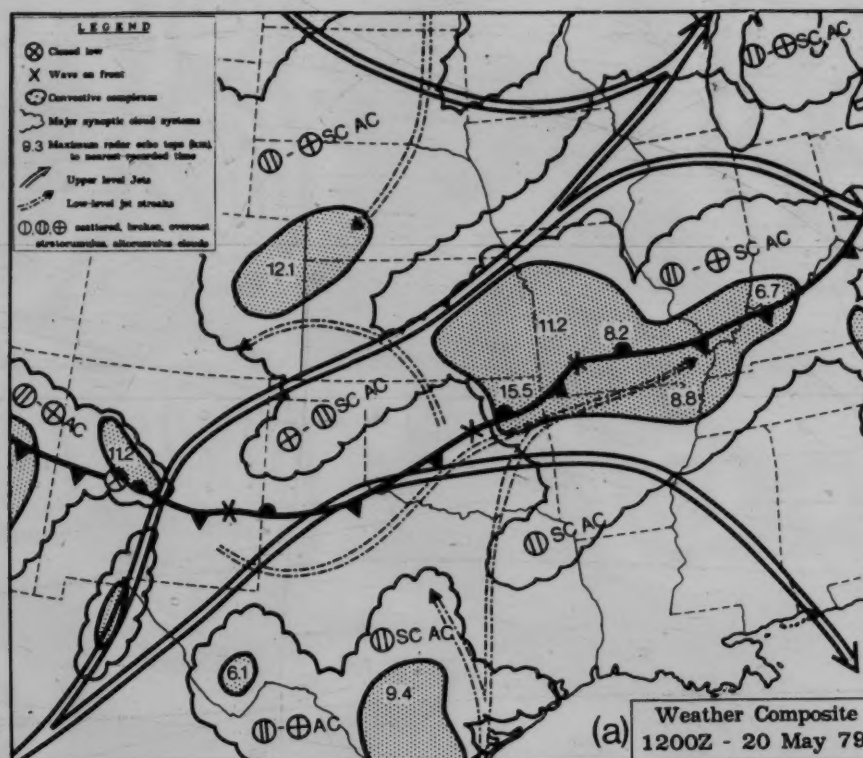


FIGURE 5.4: (caption next page).

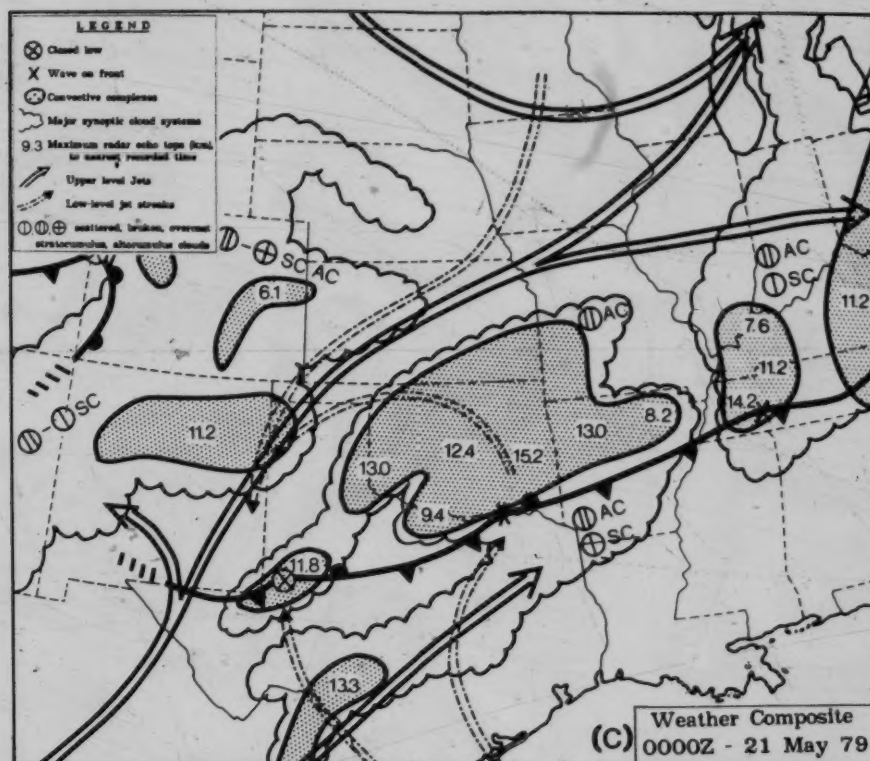


FIGURE 5.4: Schematic composites of synoptic cloud systems, convective regions, surface fronts, frontal waves, lows, and major upper (300 mb) and lower (850 mb) jet streaks for (a) 1200 UTC, (b) 1830 UTC, and (c) 0000 UTC, 20-21 May, 1979. Scalloped regions encompass low- (SC) and mid-level (AC) cloudiness (broken to overcast conditions), as determined from surface observations, soundings, and satellite imagery; stippled regions outline major convective complexes, and include the maximum radar echo tops (km) at the nearest times. The double arrows indicate main 300 mb jet streams (solid) and 850 mb wind streaks (broken).

Atmospheric mass, and therefore energy, is exchanged whenever two jet streams merge or diverge. In this sense, the two main upper jet streams, labelled "N" (northern) and "S" (southern) respectively on Figures 5.1 and 5.2, are primary energy exchange mechanisms for the above vortices. The two jets appear to merge southwest of the Great

Lakes around 201200 (Figure 5.1b), suggesting that important energy exchanges take place between the two synoptic systems. Priddy and Fuelberg (1984) compared 24-hour area- and time-averaged KE budgets between the mesoscale network and synoptic scale regions of the southeast U.S. for this case. (Their synoptic scale region was comparable to the region shown in Figures 5.1, 5.2.) Among their findings were net transports of KE out of both the mesoscale network region and the larger synoptic region at jet stream levels. This transport may represent a supply of kinetic energy to the northern system.

The southern jet had a cyclonically-curved maximum as it approached southwest Texas, with several secondary westerly jet streaks branching across Oklahoma and southern Texas until 210000. At the same time, a cyclonically-curved low-level jet streak undercut these upper jets over western Kansas at 201200, and over southern Oklahoma throughout the day. According to the Beebe and Bates model (Section 1.3.3, Figure 1.3), these jet combinations would have favoured convective formations near the Colorado-Kansas border, southwest New Mexico, and southwest Oklahoma at 201200. By 210000, this synoptic dynamic support shifted mainly to the Texas Panhandle, south-central Texas, and southeast Oklahoma.

At the same time, the combination of a low centre and two jet streaks apparently converging at 850 mb over southwest Texas, provided dynamic support for another convective storm over west-central Texas. By 211200, most dynamic support had shifted to northeast Oklahoma, with renewed dynamic support for convection over southwest Texas because of an easterly cyclonic jet branch from Texas into New Mexico. Most of these inferred convective storms are apparent in Figures 5.3 and 5.4 and

the radar composites of Figures 5.6-5.16. This demonstrates that one can deduce sub-synoptic scale storm threat regions from upper jet positions and curvature alone.

The low centres at both 500 and 700 mb coincide closely with the 250 mb centre, moving northeast from the Baja at 200000 to New Mexico by 211200. A southwest flow persisted over Oklahoma at these levels because the trough, extending northeast from the low centre, remained northwest of the state. The 700 mb trough eventually moved into Oklahoma late in the period, but it remained northwest of the major convective complex. The author's Alberta experience with 500 mb pattern recognition (e.g., Appendix Figure A1.2) would suggest a convective storm threat region covering New Mexico and western Texas at 201200, shifting into southwest Oklahoma by 210000, commensurate with general storm happenings this day.

The pattern at 850 mb, which was below convective cloud bases in this case, differed from those of the higher levels. The main low centre, with a trough extending northeast, was over New Mexico at 200000, but moved southeast to southern Texas by 211200, with the trough moving down across central Oklahoma. A southerly low-level flow was maintained south of the trough line, and it continued to draw moist air off the Gulf of Mexico. This trough remained in the vicinity of the major convective complexes over Oklahoma throughout the storm period. The juxtaposition of the front and convective storms allowed warm moist Gulf air to be pumped into the south side of the storms, while feeding cooler air in a light northeast flow into their north side. On the other hand, the surface trough and cold front moved just south of the storm formation region early in the period, so that most of these storms

formed north of the surface front in the so-called cold airmass. The mesoscale summary will describe these events in more detail.

5.1.2 Synoptic Thermodynamics

An interesting feature occurs in the thermodynamics of this situation. The cyclonic vortex over northwest Mexico in Figure 5.1, contains a cold core which is evident at all levels. During the pre-storm period, temperatures at 700 and 500 mb over the low centre are some 10 degrees lower than at the same levels over Oklahoma and Texas. This low maintains a southwest flow over Oklahoma and Texas while approaching the region, while the isotherm pattern (Figures 5.1-5.2) suggests "warm" advection at all upper levels throughout the storm period. Nevertheless, the air below 500 mb cools by 10°C in 24 hours (Oklahoma City, 200000 - 210000).

As pointed out in Section 4.5.3, this cooling resulted from ascent and adiabatic cooling of the airmass, which clearly exceeded any warming (see Figure 4.7 again) due to horizontal advection. Even without any mesoscale analyses, Figure 5.1 shows that this cooling and ascent had commenced prior to 201200, some six hours or more before any significant convection over Oklahoma. The first radar echoes were observed at 201830 (July and Turner, 1981). There is little doubt that this ascent was associated with the dominant synoptic disturbance approaching from the Baja Peninsula. It appears that the slow southward propagation of the cold front over the southeast U.S., against a push in the opposite direction from warm advection at all levels, was directly related to, and perhaps caused by, the ascent induced by synoptic processes. The

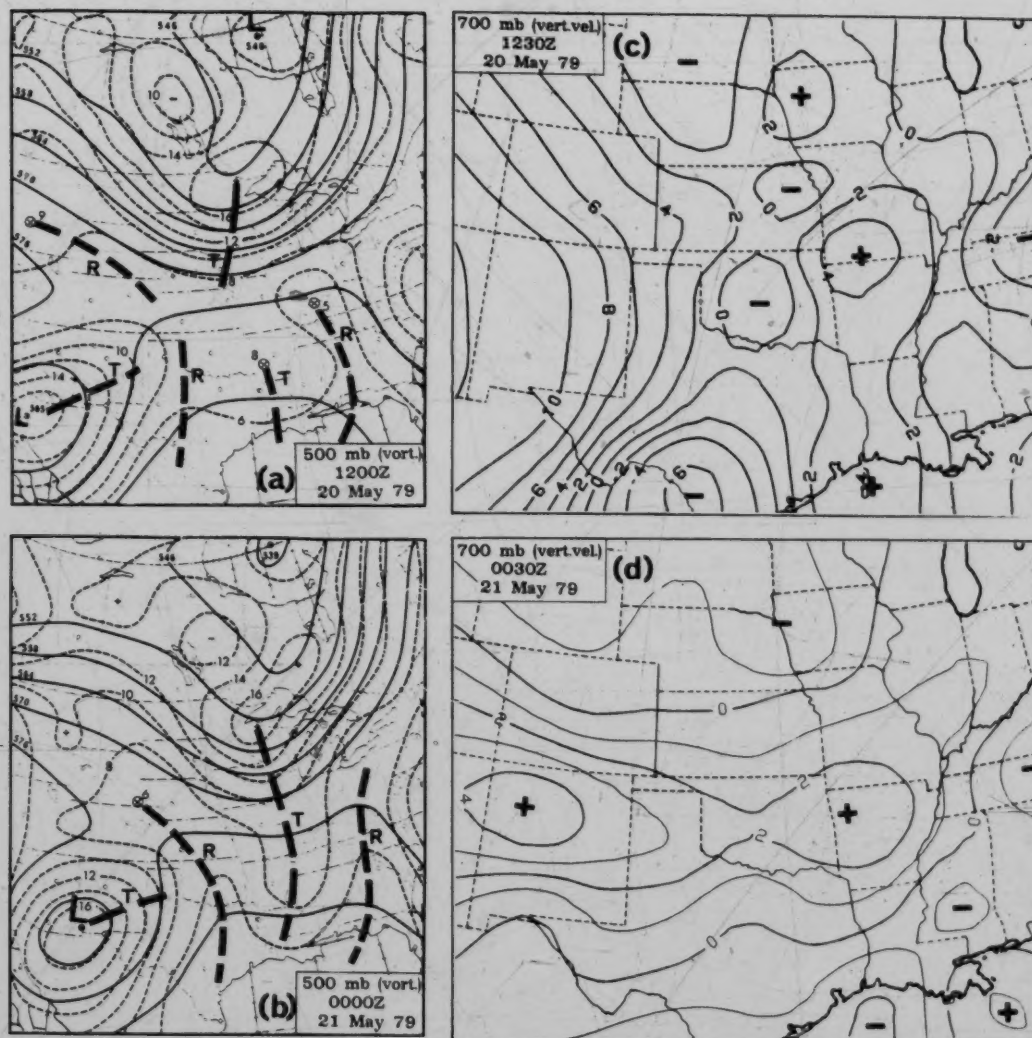
mesoscale analyses of Section 5.2 will therefore concentrate on resolving dynamic features originating from southwest of Oklahoma. These will also attempt to reveal the spatial and temporal scales of motion involved.

5.1.3 Synoptic Cloud Structures

Two main jet streams have been identified over the eastern portion of the continent (Figures 5.1-5.2). We also find two main synoptic cloud systems related to these jets (Figure 5.3). One of these cloud systems lies mainly north of the Great Lakes, the other extends southwest from the Great Lakes into Mexico. Prior to 201200, five convective complexes already existed at the start of the AVE-V experiment, apparently survivors of storms initiated the previous day. These five complexes virtually surrounded Oklahoma (Figure 5.3), but they remained distant from the SESAME mesoscale network. They will therefore be mentioned only in connection with the prevailing synoptic conditions. One of these was the Colorado storm, supported earlier by the synoptic scale upper trough and an easterly upslope flow at low levels (Figure 5.1). It was disintegrating as the trough overtook it and subsidence gradually invaded Colorado. For example, storm tops which exceeded 12 km at 201135, dropped below 8 km and continued to decline after the major Oklahoma storm was initiated at 201830 (Figure 5.4). A second complex was still quite active along the eastern portion of the cold front over northeast Oklahoma at 201135. It persisted throughout its second day, while moving into central Kentucky and Tennessee, some 900-1000 km from the SESAME mesoscale network by late

afternoon (Figure 5.4b). The third and fourth complexes over southeast and southwest Texas were in advanced states of decay by 201135, degenerating to little more than scattered showers by late morning. A fifth weak complex over central New Mexico appeared to gain support during the day as a result of the approaching major synoptic disturbance. This latter feature caused a gradual increase in cyclonic vorticity (Figure 5.5) and related ascent (Figures 5.6-5.13) over the region throughout the period.

Jet streak clouds, as described by Weber and Wilderotter (1981), can be identified in the satellite imagery (Figure 5.3), and are associated with the main and secondary jets observed on the 250 mb analyses of Figures 5.1-5.2. In particular, the southwesterly jet crossing extreme southeast New Mexico at 201200 (Figures 5.1, 5.4) has a band of mid- and high-level cloud (AC and CS) on its southeast side (Figures 5.3, 5.4). This band of cloud slowly tracked northeast with the jet core winds to extreme southwest Oklahoma, where the AVE-V storm complex formed at about 201830. A later mesoscale analysis (Appendix IVb) shows that the jet core has a maximum over south-central Oklahoma, where it also exhibits anticyclonic curvature. The right entrance of this jet maximum is extreme southwest Oklahoma, where the most severe (AVE-V) complex formed (Figure 5.4b). This is in agreement with the Beebe and Bates (1955) model (Section 1.3.3). The favourable upper level dynamics, coupled with surface frontal convergence of moist air in a low-level southerly flow, provided support for the formation of the AVE-V convective storm.



two methods gave similar but not identical results, and neither technique could be clearly judged superior from the inter-comparison alone, so that either or both techniques could conceivably be inadequate for the purpose. However, the adiabatic vertical motion fields did correspond more closely to the synoptic and sub-synoptic scale organization of clouds as seen in the GOES satellite imagery.

An additional test of the adiabatic technique is to see how well the vertical velocity field correlates with the synoptic scale dynamic features. Only data from the synoptic rawinsonde stations (with 370 km spatial resolution) are included for this comparison, while taking advantage of the 3-hour temporal resolution of the soundings to produce vertical velocity analyses at 201230 (using 1100 and 1400 UTC soundings) and 210030 (using 2300 and 0200 UTC soundings). In the absence of direct computations, a forecaster often infers the vertical motion field from other synoptic features, such as the 500 mb vorticity advection (with positive vorticity advection (PVA) implying ascent, and NVA subsidence). Miller (1972) lists PVA at 500 mb as "the most important parameter for severe weather forecasting".

In Figure 5.5 therefore, we compare the computed adiabatic vertical velocity fields with the synoptic patterns of 500 mb PVA and trough-ridge positions at 201200 and 210000. There is good agreement between the large-scale features at both times, if one allows for slight variations in position. For example, the north-south line of maximum vertical velocity east of Texas-Oklahoma at 201230 can be associated with a shortwave trough and a corresponding band of weak PVA at 201200. Likewise, the north-south line of minimum vertical velocity through central Oklahoma has a similar band of weak NVA east of a shortwave

ridge. The broader region of strong ascent generally west of Texas relates well to the PVA over those regions, with the major synoptic vortex approaching from northern Mexico. Similar agreement is evident at 210000. The fact that the synoptic scale vorticity advection is quite weak in this instance, is consistent with the severe storm findings of Hales (1979) and Maddox and Doswell (1982).

In general, the vertical motion fields of Figure 5.5 also relate better to the cloud-free regions, and both mid-level clouds and large convective complexes, as summarized in Figure 5.4, than do the synoptic vorticity patterns. For example, the horizontal extent and the orientation of the relatively cloud-free regions over Oklahoma, southwest of the Great Lakes, and over northeast Mexico at 201200, and over southern and southwest Texas and north of Kansas at 210000, correlate well with the regions of subsidence in Figure 5.5. Centres of ascent over southwest Missouri, and maxima bulging into New Mexico and western Kansas at 201230, correspond to major convective storm clusters already in progress in those vicinities. By 210000, the major convective complexes lie in a broad west-east line (Figure 5.3 and 5.4c), quite similar to the synoptic scale vertical motion pattern (Figure 5.5d), with major storm complexes corresponding to the main centres of ascent.

The foregoing discussion confirmed that there is a close correspondence between patterns of adiabatic vertical motion and the synoptic features upstream (southwest) of the convective complexes. The next step is to examine mesoscale responses to these features over the AVE-V mesoscale network, investigate apparent source regions southwest of Oklahoma, and in doing so, test the spatial and temporal continuity

of the computed mesoscale vertical velocity fields.

5.2 MESOSCALE VERTICAL MOTION PATTERNS

In this section, analyses are dubbed sub-synoptic or mesoscale according to the resolution of the available data as discussed in Section 4.1, and according to the scales of motion defined in Table 1.1. Vertical motion fields at 800, 700, and 600 mb are compared with fields of surface convergence and observed convective complexes, using GOES visible images and radar composites (Figures 5.6-5.16). The radar composites are reproductions from July and Turner (1981). The hourly surface analyses referenced here appear in Appendix IV.

It is important to remember that the 80-km spatial resolution of the vertical velocity analyses is only over western Oklahoma (mesoscale network, Figure 4.2). The spacing increases to 370 km over the remaining synoptic network (average values given in Section 4.1.1), while all of the computed results are three-hour averages. Because of this, at best, one can only retain spatial and temporal continuity from map to map for features of horizontal scale size 160 km or more over the mesoscale network. Over the synoptic network, the minimum resolvable feature is 740 km. For the surface data, with average data spacing of 130 km throughout the U.S. (Figure 4.3), only features of 260 km are resolved.

In the following mesoscale summary of vertical motion patterns and related convective features, the descriptions are broken down into four periods covering the pre-storm, storm initiation, mature storm, and

decaying storm stages. Frequent references will be made to the major synoptic features just discussed, and to sub-synoptic scale features of the hourly surface analyses which appear in Appendix IVa. First, recall that there is a large cyclonic vortex approaching the region from northwest Mexico (Figure 5.1). Downstream from this synoptic system, and in response to the resulting PVA, low surface pressures over New Mexico and western Texas remain a persistent feature during AVE-V (Appendix IVa). The only large region of ascent (Figure 5.6) is generally west of Texas, is directly associated with the PVA and surface low pressure system, and it persists throughout the period. Other vertical motion features are primarily mesoscale in both time and space. A sharply-defined surface front extends east-northeast from the low pressure centre over New Mexico (Appendix IVa).

We note in Figure 5.4a, that one of the upper jets crosses the surface front over central New Mexico, coincident with the main large centres of both low surface pressure and mid-level ascent throughout the pre-storm period. The hourly surface analyses (Appendix IVa) suggest that various (mesoscale) frontal waves may be emanating from near this intersection region and moving northeast along the front. The surface data alone, are of too coarse a resolution (130 km) to verify the exact source regions of these mesoscale frontal waves and surface pressure centres without ambiguity. The other upper jet intersects the front near extreme southwest Oklahoma at 201200, where the major severe storm later forms. The pressure and vertical motion fields are therefore now examined together, in an attempt to determine if these two regions around central New Mexico and northwest Texas are sources for both the frontal waves and the mesoscale centres of ascent.

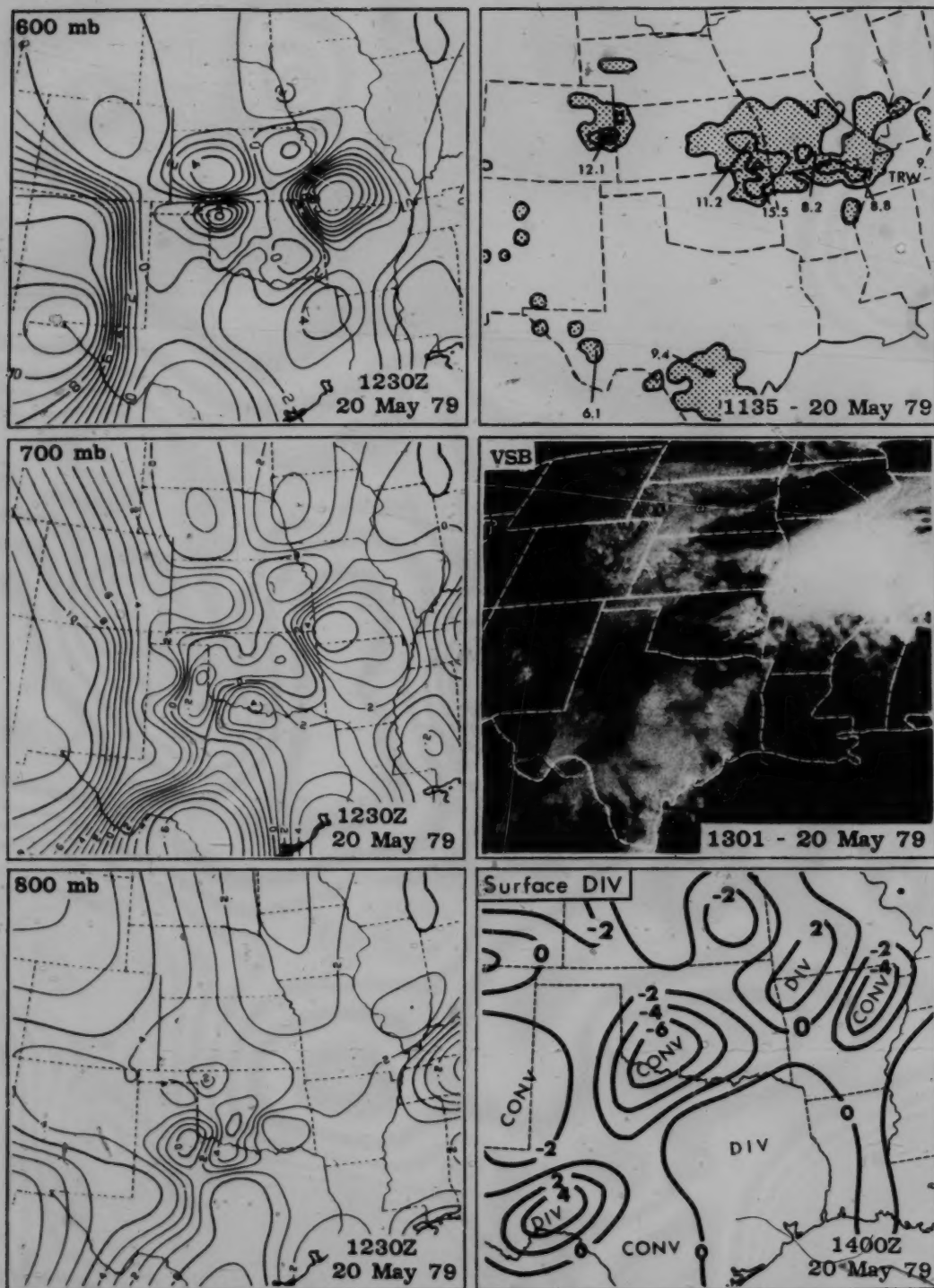


FIGURE 5.6: 1230UTC vertical velocity(cm s^{-1}), radar composite, and GOES satellite imagery, 1135-1301 UTC, and 1400 UTC surface divergence (after Sikdar and Fox, 1983), 20 May, 1979. NOTE: most GOES images here and in subsequent figures were re-gridded to correct errors on original pictures.

5.2.1 The Pre-Storm Period

At 201230 (Figure 5.6), there remains significant convection over eastern Colorado (decaying complex), northeast Oklahoma (older, but re-intensifying complex), southeast Texas (decaying complex), southwest Texas (transient scattered cells), and New Mexico (scattered cells forming). Vertical velocity peaks associated with each of these regions can be noted, although they are displaced slightly eastward from the Colorado and southeast Texas complexes. This apparent displacement may result from the poor spatial resolution of soundings outside the Oklahoma mesoscale network, but it may also reflect the decaying state of the two complexes, since synoptic support had over-taken both (note the shortwave ridge position in Figure 5.5a).

The relatively cloud-free band running from southwest to northeast through Oklahoma is related to a region of subsidence, especially at 700 mb (Figure 5.6). The clouds northwest of this clear region (from the Texas Panhandle to western Kansas on the satellite image), are generally low- to mid-level according to surface observations. These clouds are seen to be partly orographically-induced when one notes the topography (Figure 4.1) and the northeast flow at low levels (Figure 5.1). They also correspond well with the ascent of $2-4 \text{ cm s}^{-1}$ at 800-700 mb over this region (Figure 5.6). This is not a conflicting argument, since the adiabatic computations yield the net vertical motion from all factors external to the convective storm.

One other noteworthy feature of Figure 5.6, is the close agreement between the general band of 800 mb ascent at 201230 and the alignment of the surface front (Figure 5.4a). The position and alignment of the

front is not so evident in the satellite imagery.

Some early mesoscale features of the vertical motion field (at 800 and 700 mb) do not appear to be related to any particular convective cloud structures in the satellite or radar images. For example, at 201230, the maximum over southern Oklahoma (values exceeding 4 cm s^{-1} at 700 mb) with a secondary bulge of ascent to near Midland (MAF), appears at all three levels (Figure 5.6), but it has no associated convective clouds at the time. However, this feature persists and intensifies on subsequent analyses, and significant storms do form in its vicinity several hours later (Figures 5.7-5.10).

The line connecting these features appears to be related to the position of the surface front (Figure 5.4). The wind field associated with the surface pressure pattern continued to draw moisture below 850 mb up from the Gulf of Mexico, while the front remained quasi-stationary through Texas and Oklahoma (see the hourly surface maps in Appendix IVa). Meanwhile, a large cell of high pressure was moving southward from the northern states, pushing a colder airmass up against the Rockies.

Sikdar and Fox (1983) were able to show that significant surface wind convergence existed over southwest Oklahoma for this case at least as early as 201400. Their analysis is reproduced in Figure 5.6f. The broad features of this convergence field at 201400 agrees with the 800 mb vertical velocity analysis at 201230. The smaller mesoscale differences occur at least partly because of the combination of different data resolution used, different times and time resolution, and because two different data bases, one being wind data, the other

thermodynamic, at two different levels are being used. The location, size, and shape of the region of convergence over the AVE-V network at 201400 (Figure 5.6f) is also remarkably similar to that of the convective complex viewed by the GOES satellite six hours later at 202000 (Figure 5.12). The complex shape is the result of a number of individual thunderstorms and anvil clouds. The time lag is not unusual, considering that the Sikdar and Fox analysis is based on the sub-synoptic scale surface data, and represents an average or composite of the various smaller scale features over the region.

Pre-storm surface convergence at least 90 minutes prior to first radar echoes has also been reported by Lewis et al. (1974), Sanders and Paine (1975), and Ogura and Chen (1977), among others. The importance of this convergence, which appears to be a pre-condition for severe storms, should not be over-looked when considering storm causes. Regrettably, most diagnostic studies have concentrated their investigations on the (visible) mature storm period, and tend to ignore or underestimate the significance of the pre-storm dynamics.

By 201530 (Figure 5.7), positive values of vertical velocity continued over Colorado and west Kansas, but the centre of ascent had almost disappeared while convective activity continued to die down. The region of ascent east of Oklahoma had also diminished, and convective activity had died down there for a few hours. Thundershowers were no longer being reported over southeast Texas, where there was now just weak ascent to support the band of mid-level clouds and the few remaining showers. The large region of ascent west of Texas relates to the scattered thundershowers, hail, and heavy rain occurring there.

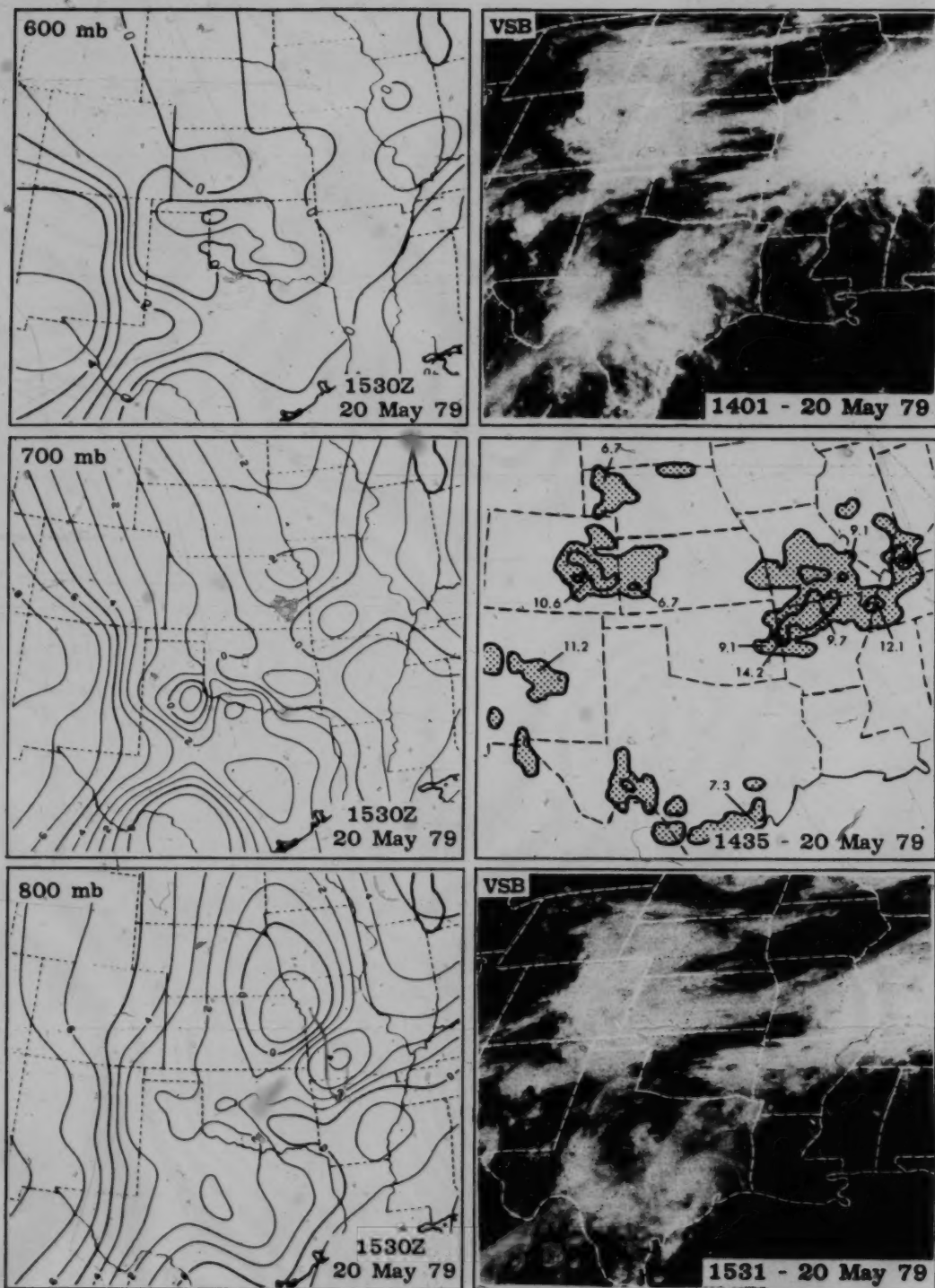


FIGURE 5.7: 1530 UTC vertical velocity (cm s^{-1}), radar composite, and satellite image for 1401-1531 UTC, 20 May, 1979.

Only a small region of weak subsidence persisted at 700 mb southwest of Oklahoma, now cut off from the subsidence over eastern portions of the state. Virtually all of Oklahoma and Texas had ascent at 800 mb, despite surface warming which, in the adiabatic method, tends to give subsidence at low levels. In spite of this predominant ascent at 201530, there was no significant convection over Oklahoma or Texas, other than a few dissipating showers over southern Texas. The southeast-northwest ridge of maximum ascent over extreme southwest Oklahoma had now extended to 600 mb, and it may have been related to a wave on the surface front over that region. Important to the discussion here, is the fact that the location and alignment of this pre-storm ascent corresponds approximately to that of initial convection which occurred three hours later, and eventually formed the major AVE-V convective complex.

By 201700, convection was at its lowest ebb for the day over the south-central U.S. (Figure 5.8). The first sign of renewed activity, however, occurs at 201740 with radar echoes of a thunderstorm over south-central Texas. This convection may have been related to a second ridge of ascent aligned west-east across central Texas (Figure 5.7, for 201530), with a distinct bulge remaining near Midland (MAF), although this portion of the vertical motion analysis is of course based on synoptic scale data only. The weather remained quiet over Oklahoma and northern Texas for another hour.

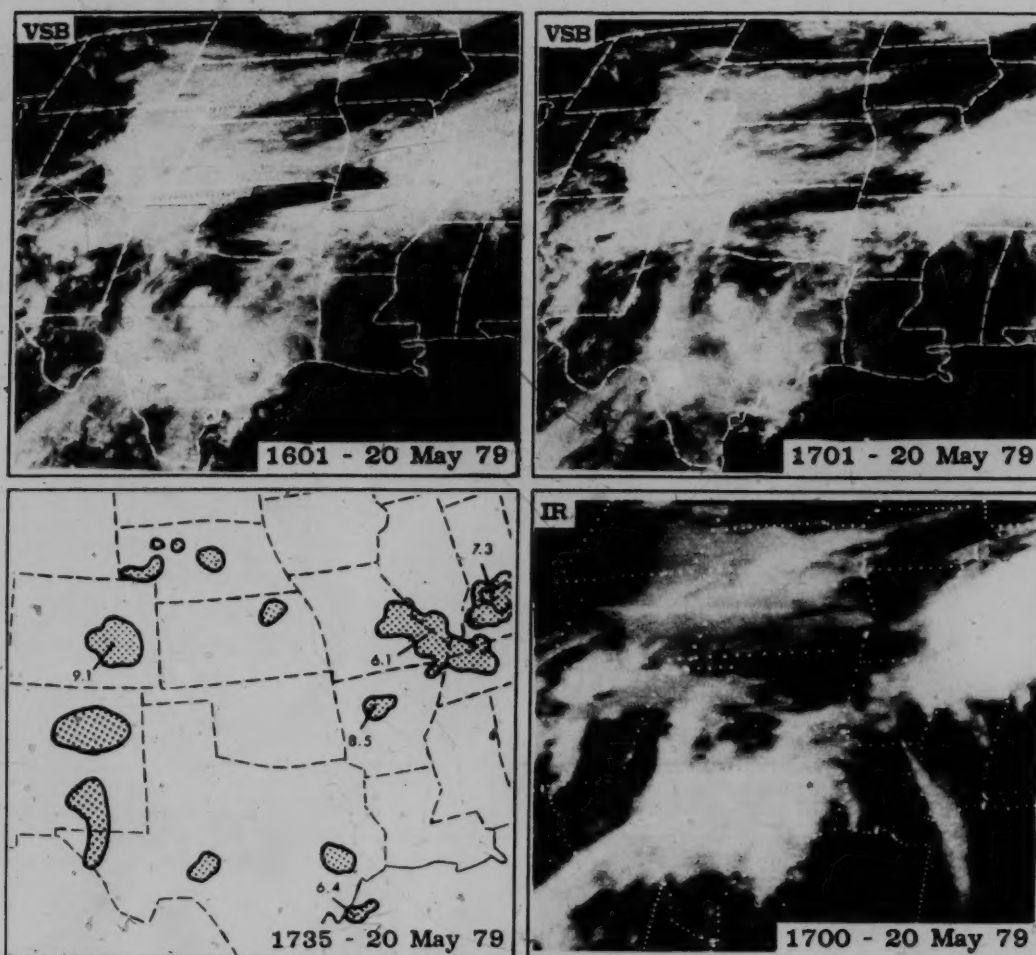


FIGURE 5.8: Radar composite and various satellite imagery for 1601-1735 UTC, 20 May, 1979.

5.2.2 Storm Initiation Stage

The surface front and trough over eastern Oklahoma started to move southward between 201700 and 201800 (Appendix IVa), shortly before the AVE-V SSCC formed. Waves were evident along the front over southeast Missouri, southwest Oklahoma, and west Texas, where the major centres of convective activity existed or commenced within an hour. The eastern portion of the front accelerated southward between 201800 and 201900,

after which the whole front showed noticeable southward propagation.

Ascent dominated both the 800 and 700 mb vertical motion fields at 201830 (Figure 5.9), except for subsidence east of Texas and north of Kansas, where skies were generally clear or scattered. A significant centre of ascent appears near Midland (MAF), west Texas at about this time, with values exceeding 5 cm s^{-1} at both 800 and 700 mb. Convective activity seemed to explode around 201825, first with thundershowers and hail reported southwest of Midland (at Pecos, Texas), corresponding with the above centre of ascent. A severe thunderstorm warning was subsequently issued for this region at 201840. Within minutes, a second major storm started to develop over extreme southwest Oklahoma. (This is the storm being referred to as the AVE-V complex.) At the same time, new storms also formed over the Texas Panhandle. These formed two lines of intense thunderstorms, oriented southeast and south-southwest from the Panhandle storm area, one connecting to the AVE-V storm area, the other to the Midland storm. Shortly afterward, other severe storm cells formed across southern Oklahoma, becoming part of the AVE-V SSCC. Other less organized thunderstorms evolved across central Texas east of the Midland storm.

These storm patterns correspond closely to the mesoscale centres of ascent (Figure 5.9), allowing for some differences in positions of the related features. The AVE-V SSCC, for example, grew rapidly to almost twice the size of the whole mesoscale network, and one must expect some aliasing of the vertical motion pattern near and outside the mesoscale network domain. The correspondence between the storms (Figure 5.4b), surface pressure patterns and frontal features (Appendix IVa), and vertical motion patterns (Figure 5.9) is also apparent.

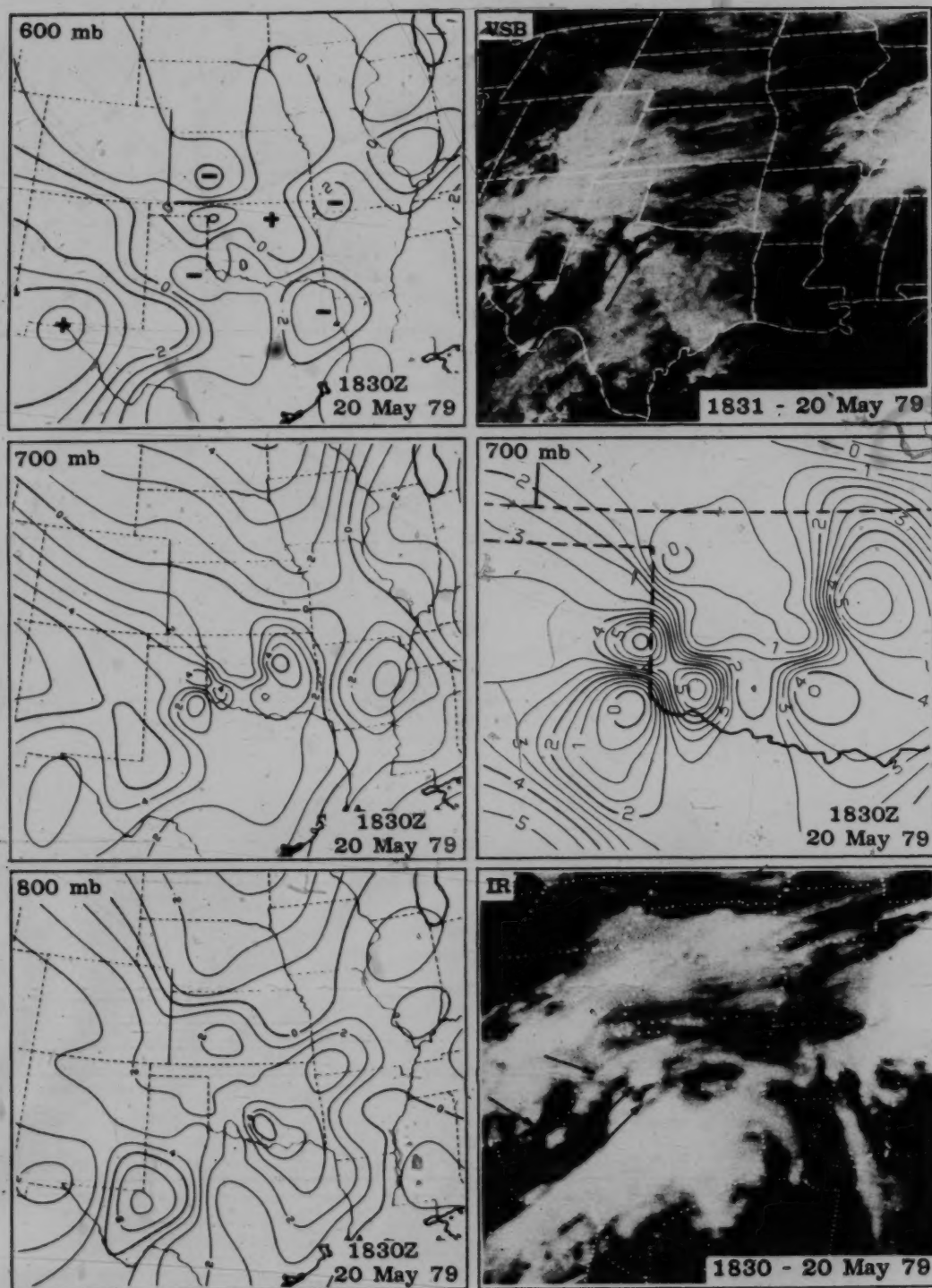


FIGURE 5.9: 1830UTC Vertical velocity (cm s^{-1}), including fine resolution 700 mb, and satellite imagery (visible/infrared) at AVE-V storm formation time, 1830 UTC, 20 May, 1979.

The major surface low centres over southeast and west-central New Mexico at 201800, and connected by a frontal trough, corresponded to the major centres of ascent throughout the period. The surface front extended northeast from the Midland low pressure centre throughout most of the AVE-V period, and at 201800, a secondary low centre and frontal wave were apparent (Appendix IVa) just southeast from where the AVE-V storm formed a short time later. The 130-km spatial resolution of the surface pressure data was not fine enough to categorically resolve the secondary low centre any earlier than 201800. This might have helped to establish the exact origins of the mesoscale motion.

A second important observation from the surface analyses is that most of the major storms of this case, with the notable exception of the Midland storm, formed north of the surface position of the front, in the cooler air. The 800 mb pattern of ascent at 201830 still closely resembles the surface frontal position across Texas and Oklahoma. Related to this, the centres of ascent tend not to be exactly coincident with secondary low pressure centres, such as the one evident over southern Oklahoma between 201800 and 202300 (see Appendix IVa). Instead, the ascent centres are located further north with height, that is, further into the colder airmass. This suggests a tilted mesoscale circulation, with the 800 mb centres generally south of those at 700 mb, perhaps reflecting a streamwise tilt of the inflow and outflow of these convective complexes. This would be analogous to the (smaller) tilted updrafts of single convective cells, and may be similarly helpful to the overall growth of the complex. That is, at the mesoscale, which we observe in these vertical motion patterns, sensible and latent heat, whose source is at lower levels and mainly south of the storm, is

released as kinetic energy at higher levels downstream, allowing continued energy conversion (and new storm cells) at the lower levels.

Focussing on the AVE-V SSCC, one can discern the beginnings of at least two smaller MCCs just southwest of Oklahoma on the GOES imagery of 201830 (indicated by arrows). The first radar echoes from these two storm complexes were reported at 201835 (radar composite not available), and minutes later, a third one was observed over south-central Oklahoma. These three distinct storm centres are still apparent on the radar composite of 201935 (Figure 5.10). Later GOES images (from 202001 on, Figure 5.10) show one large sub-synoptic scale convective complex covering all of Oklahoma. Three cells of 700 mb ascent are similarly resolved over the region at 201830 (see inset of finer resolution in Figure 5.9), but as in the case of the GOES cloud picture, the two closer cells appear to be merged into one by 202130 (Figure 5.11).

These storms intensified so rapidly that tornado watches were issued for most of southern Oklahoma at 201845, and for the Midland region at 201900. These watches were confirmed and updated to warnings following tornado sightings and hook echo observations over southwest Oklahoma within or just south of the mesoscale network at 201930, 201955, and at 202040, along the ridge of 700 mb ascent of Figure 5.11.

Other aspects of the vertical velocity field at 201830 also correspond well with observed mesoscale cloud features. This includes the cloud-free regions southeast and southwest of the Oklahoma complexes, and severe, although less well organized storms over northern parts of New Mexico. Most storm cells throughout the three states moved northeast at 40 km hr^{-1} , with the larger storm complexes having a more

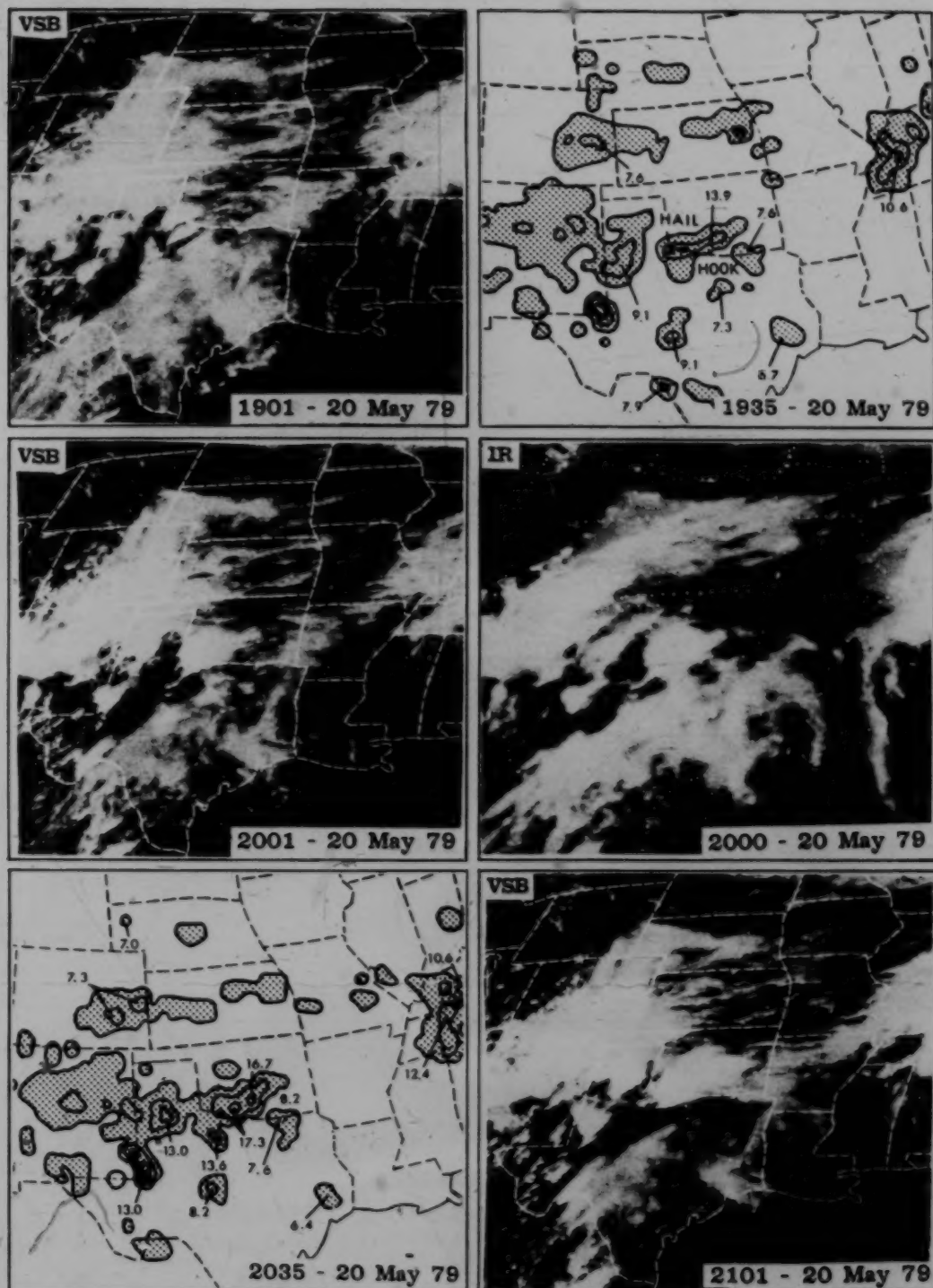


FIGURE 5.10: Radar composites and satellite imagery during the storm development stage, 1901-2101 UTC 20 May, 1979.

easterly component. Further afield, the Colorado storm complex continued to dissipate, with no discernible mesoscale centre of ascent remaining at 201830. According to the radar composites, the eastern storm complex became reactivated after 201830, but it had moved well into Kentucky-Tennessee, coincident with a centre of ascent over that region.

5.2.3 The Mature Storm Period

The AVE-V complex reached severe storm status less than an hour after the first radar echoes. Radar echo tops exceeding 15 km were continually reported between 201930 and 210430. At the peak intensity, tops to almost 20 km were attained some 55 km northwest of Wichita Falls (202005), just within the southern portion of the mesoscale network. A tornado was sighted in that same vicinity at 202040. Two or three MCCs (within the AVE-V SSCC) are evident in the radar composites for 201935 and 202035, but only one large SSCC can be distinguished in the GOES visible imagery after 202000 (Figure 5.12), the cirrus cloud blow-off covering most of the state and obscuring the smaller MCCs.

Four main centres of ascent are maintained on the 700 mb vertical velocity analyses of 202130 (Figure 5.11). These are over eastern Oklahoma, the eastern part of the Texas Panhandle, near Midland, and over northern New Mexico. While the locations of ascent centres (to mesoscale resolution) would be unknown since they are outside the mesoscale network, the regions of ascent do correspond to four main regions of convective activity as indicated by the satellite imagery and radar composites.

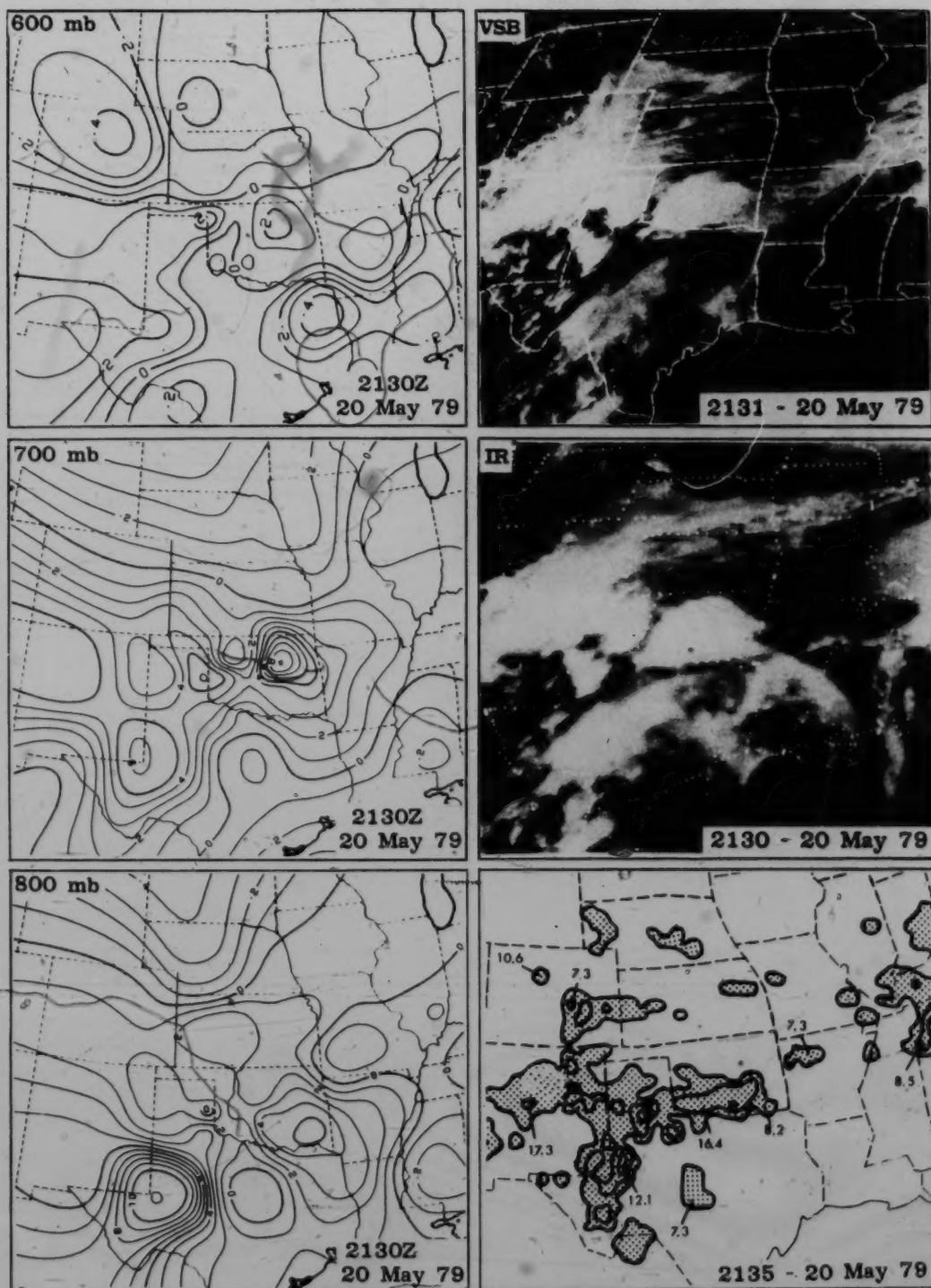


FIGURE 5.11: 2130 UTC vertical velocity (cm s^{-1}), radar composite, and satellite imagery (visible/infrared) at the mature storm stage, 20 May, 1979.

When considered together with the 800 and 600 mb vertical motion patterns, the Oklahoma and Midland ascent centres are the most intense at 202130, as are the convective complexes in those two regions. In conjunction with the four main centres of environmental ascent at this time, very heavy thunderstorms with hail (TRW++A) were reported at Albuquerque (ABQ) at 202155, Midland (MAF) at 202258, near Childress, Texas (CDS) at 202256, Oklahoma City (OKC) at 202254, and McAlester (east-central Oklahoma) at 202257. Tornadoes were also observed on the south edge of the western storm cell of the AVE-V complex, and near OKC at 202254 (Figure 5.12). In addition, ascent had replaced subsidence over Arkansas by this time, where a line of showers was detected over the northwest portion of that state at 202135, reaching intense thunderstorm (TRW++) status by 202235. Elsewhere, convective activity had dissipated over the greater southeast portion of Texas, over the El Paso, west Texas region, and over the west side of the Texas Panhandle. All three of these areas exhibited negative or minimum values of vertical velocity by 202130.

The adiabatic patterns of ascent at 600 mb also show good correspondence with the observed convective complexes at 202130. One cannot discount some error due to cloud latent heat effects, which would cause underestimates of ascent and overestimates of subsidence. However, the patterns fit well, including centres of ascent slightly downwind of (generally northeast of in this case) storms over northeast Oklahoma and the Texas Panhandle, and subsidence in the rear of the AVE-V complex over western Oklahoma and over the relatively clear eastern Texas, along with downslope subsidence in the lee of the Mexican Plateau over south-central Texas.

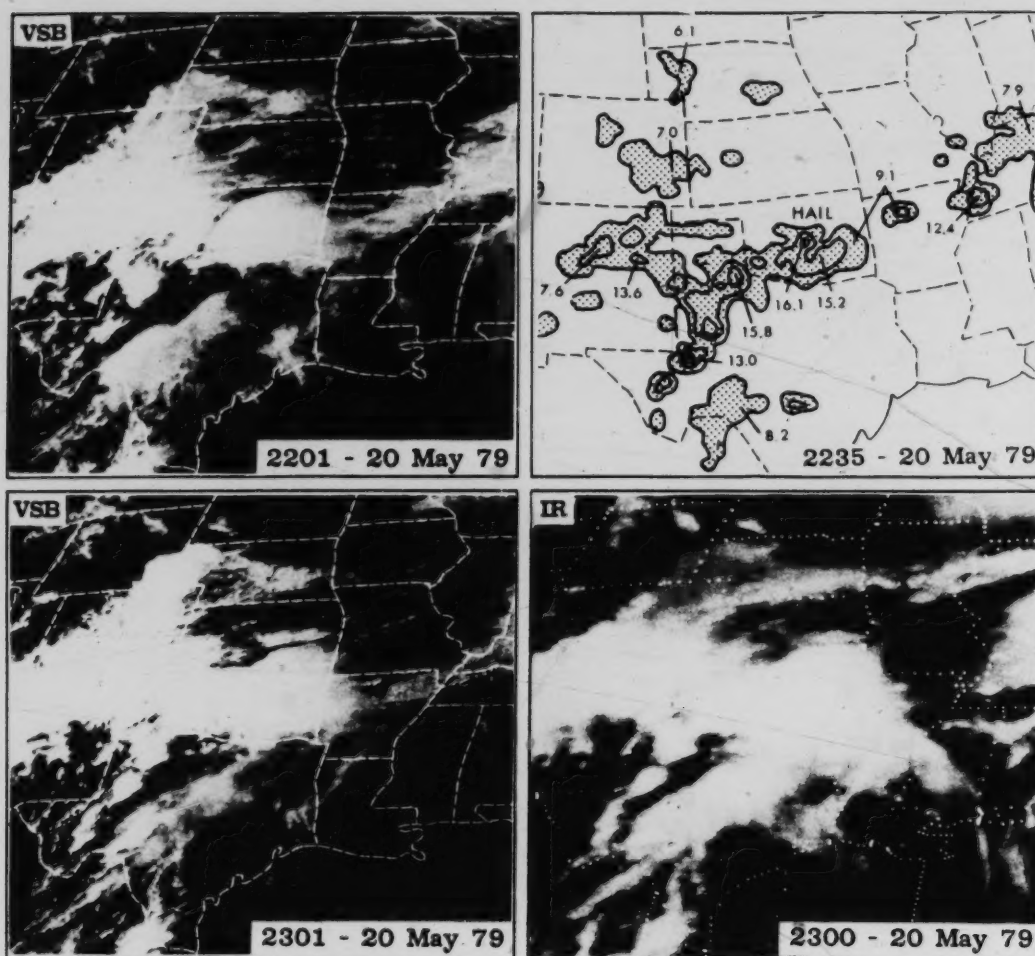


FIGURE 5.12: Radar composite and various satellite imagery (visible/infrared) during mature storm stage, 2201-2301 UTC, 20 May, 1979.

Once again, it is emphasized that the vertical motion fields are presumed to be those of the environment around storms, and not the storm scale motion itself. In other words, one should not necessarily expect strong environmental ascent above 700 mb outside of active convective storm cells at this distance from the approaching shortwave (see Figure 5.5 again).

The 800 mb pattern of ascent at 202130 again gives the impression of sub-synoptic to mesoscale low-level inflow and ascent on the south to southeast side of the convective complexes, with outflow on the north to northeast sides. The ascent centres associated with the AVE-V and Midland storms, particularly the latter, were dynamically related to the low-level southerly jets (Figure 5.4b) which were supplying moisture to these regions.

During the storm development stage, the surface front slipped further south across Oklahoma and Texas, so that by 202200, even the Midland storm was north of the surface front. Weaker storms were intensifying in the warmer airmass southwest of Midland, and over south-central Texas, near the ridge of high ascent values (Figure 5.11).

By 210030, two neighbouring thunderstorms over southwest Oklahoma, showing radar echoes of 13 and 15.5 km, were associated with a centre of ascent at 800 mb with values exceeding 4 cm s^{-1} (Figure 5.13). At 700 mb, there appeared to be two centres of ascent associated with the two cells. Also, a small region of subsidence at 600 mb over southwest Oklahoma agrees with the clear region between the two small storms (see the GOES visible image at 210001). In other words, the pattern of vertical motion over Oklahoma remained quite similar to the radar and satellite depictions of the storms. The centre of ascent previously over the Midland location, did not show up at 800 and 700 mb this time, as the associated storm had moved northeastward. One notes that at least three tornado sightings were reported 30-100 km northeast of Midland between 202345 and 210020, where previously they were being observed further southwest.

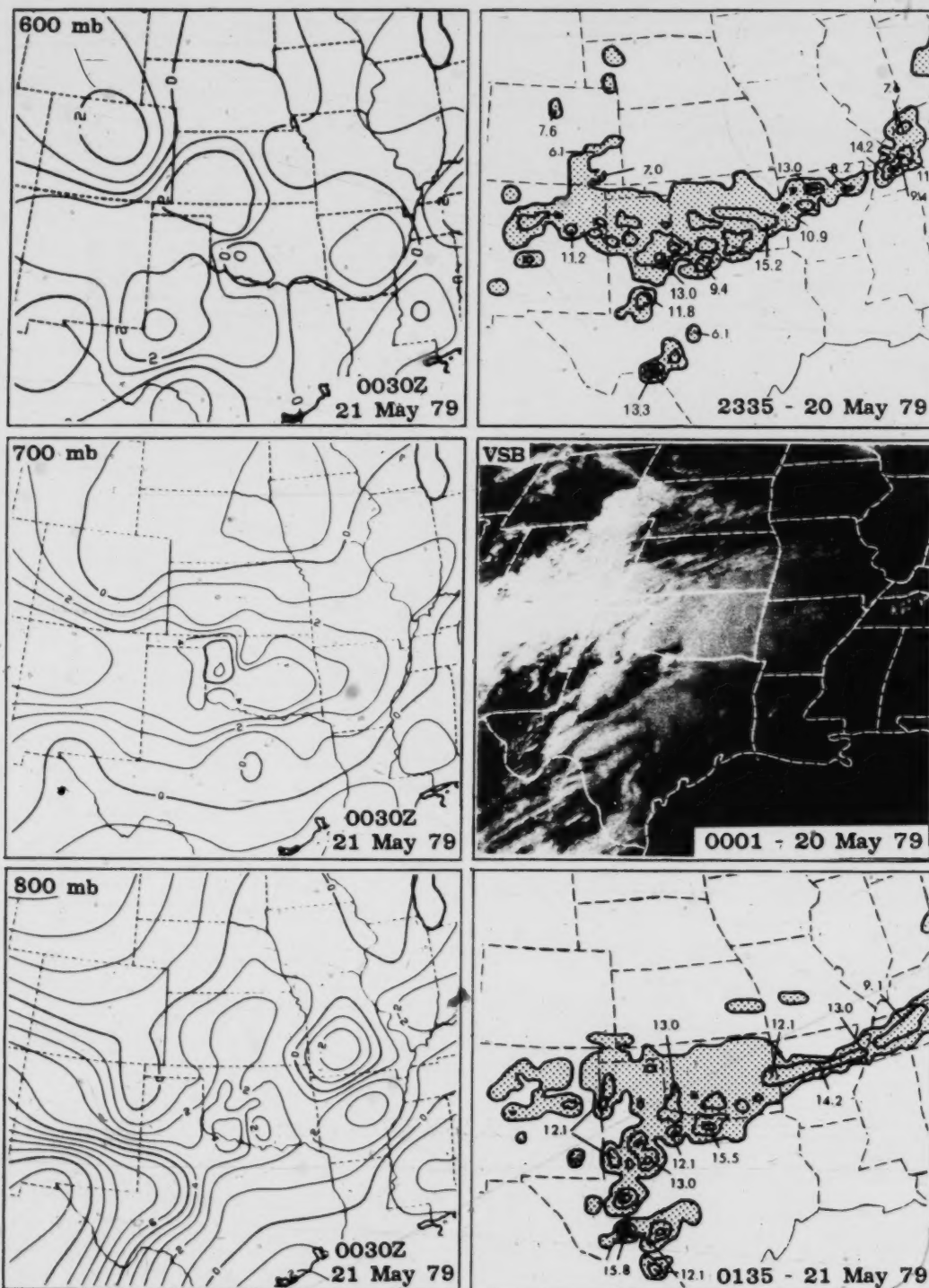


FIGURE 5.13: 0030 UTC vertical velocity (cm s^{-1}), radar composites, and satellite image during the mature storm period, 2335-0135 UTC, 20-21 May, 1979.

Low-level ascent increasing over southwest Texas was due to strong PVA as the major synoptic vortex crossed into New Mexico (Figure 5.2). This ascent persisted until after 210630, and may have played a role in the formation of new storms just east of there. Environmental subsidence above 700 mb over this new storm region may have been simply the result of persistent downslope effects in the lee of the Mexican Plateau. Apparently, any contamination of vertical velocity values by latent heat errors would not seem to be appreciably affecting either the sign of computed values or the patterns.

By 210330, the line of maximum ascent which had been over southern Oklahoma at 210030, was moving southward into northern Texas (Figure 5.14). While individual storm cells continued to move east-northeast, the sub-synoptic scale convective complex had gradually slipped southeast as well. Radar echo tops which exceeded 15 km just south of Oklahoma, correspond with centres of ascent at 800 and 700 mb. At the same time, there was a brief resurgence in surface frontal wave activity just south of this region. Within the scale of data resolution, the 800 mb pattern still corresponded very closely with the surface frontal position and frontal waves.

Meanwhile, the approaching synoptic vortex was accompanied by much drier, subsiding air from the high Mexican Plateau, so that even with ascent over the El Paso region at 210330 (Figure 5.14), the air was too dry to support new thunderstorms. At the same time of course, all storms were losing their surface heat source as the sun set.

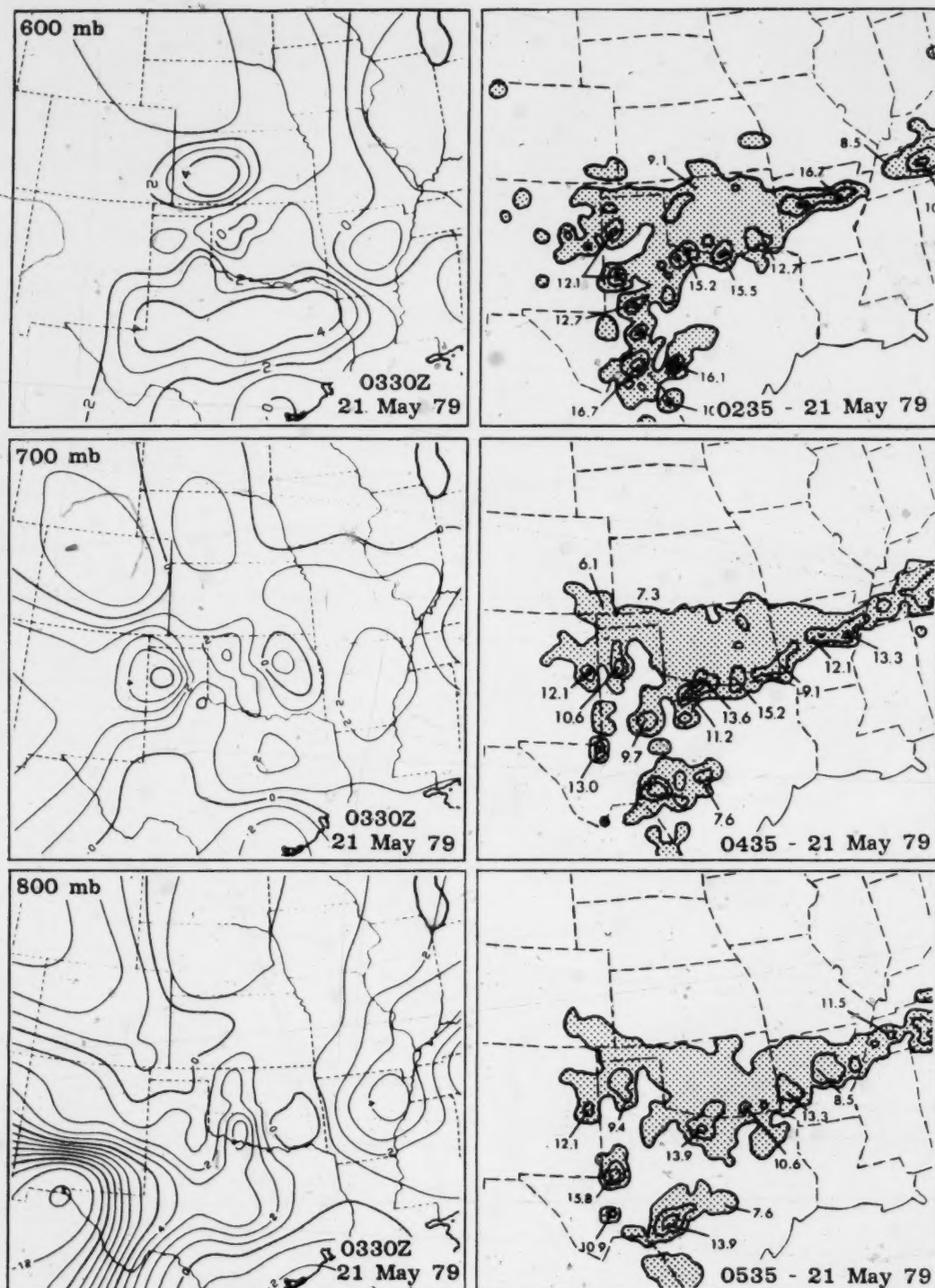


FIGURE 5.14: 0330 UTC vertical velocity (cm s^{-1}) and various radar composites, 0235-0535 UTC, 21 May, 1979.

However, over the west side of the Texas Panhandle where mid-level ascent had increased, storms from northern New Mexico and west Texas (Midland storm) had merged and intensified near Amarillo after 210235. A similar merging of the two comparable ascent centres between 202130 and 210330 is suspected, based on their history, and because a significant increase in synoptic scale vorticity was occurring there at about this time according to Figure 5.5.

5.2.4 The Decaying Storm Period

It is difficult to say when the large complexes actually started to decay. Radar echoes as high as 15 km were still reported near the southern Oklahoma-Texas border up until 210430. However, the time 210500 was chosen as the beginning of the dissipation stage for the AVE-V complex, since major radar echoes appear to have moved south of Oklahoma by 210535 (Figure 5.14), and a line of minimum vertical motion appeared across southern Oklahoma by the 210630 analysis time (Figure 5.15).

The main centres of ascent and the associated convective storms had shifted eastward from Oklahoma to Arkansas, and from west Texas to the Panhandle by 210630. A weak line of maximum vertical velocity remained over eastern Oklahoma, where widespread showers and isolated thundershowers lingered on, but were slowly dissipating. With the approaching synoptic vortex, the centre of ascent which had been near El Paso at 210330, moved over the Midland region. There, it combined with an existing source of moisture which continued to be advected up the Rio Grande and Pecos river valleys from the Gulf of Mexico (see surface

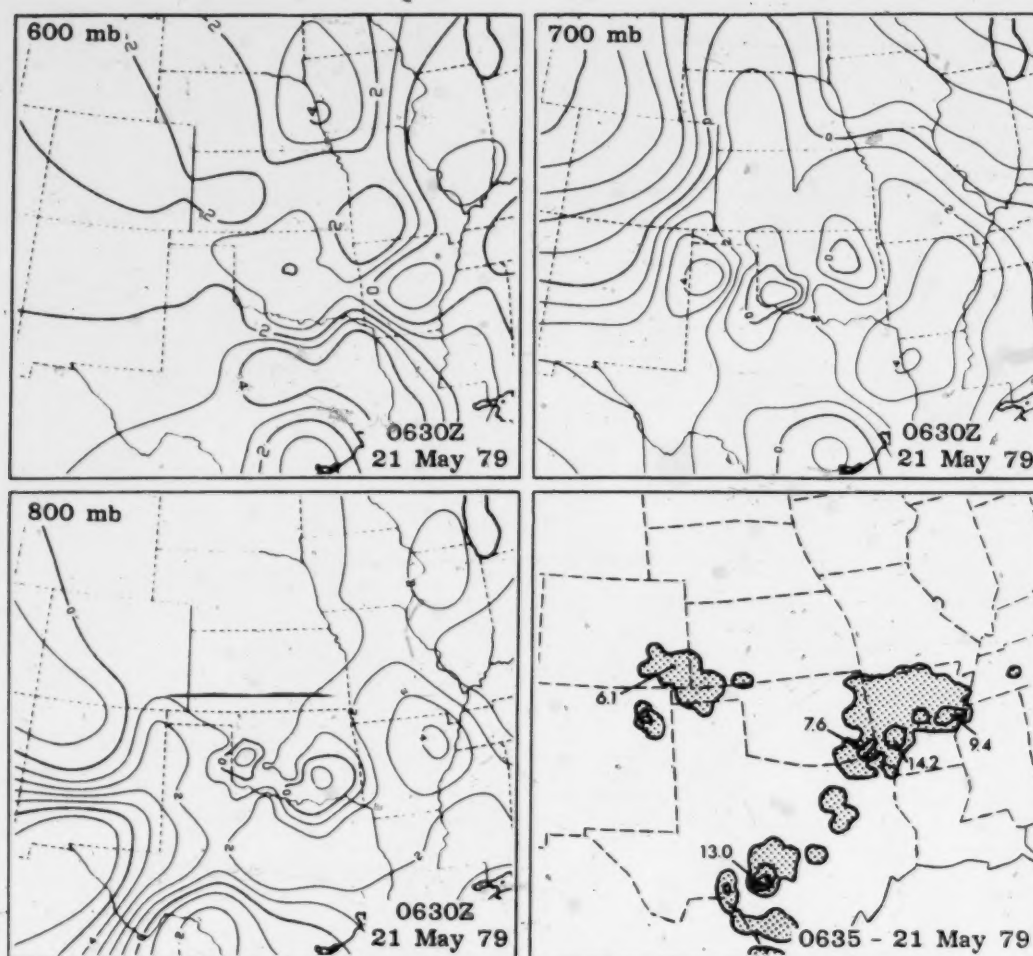


FIGURE 5.15: 0630 UTC vertical velocity (cm s^{-1}) and 0635 UTC radar composite, decaying storm period, 21 May, 1979.

analyses, Appendix IVa). This resulted in a resurgence of storms in that region, generating new storm cells beyond the AVE-V data period. Extreme western Oklahoma also experienced new storm activity after 210730, as the area of ascent from the Texas Panhandle gradually moved over that region by 210930 (Figure 5.16). The region of ascent over eastern Texas and Louisiana at 210930 also appeared to be a major factor in new storm formations there.

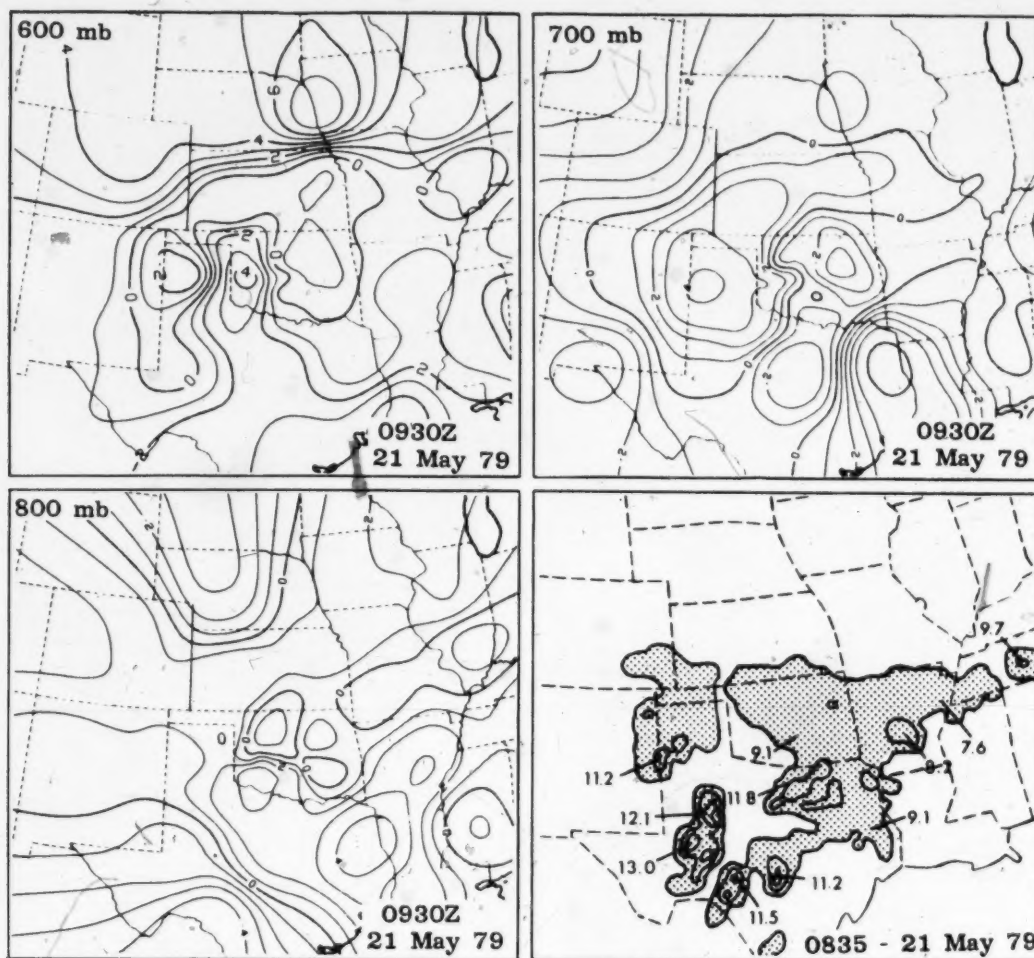


FIGURE 5.16: 0930 UTC vertical velocity (cm s^{-1}) and 0835 UTC radar composite, decaying storm period, 21 May, 1979.

5.3 CONTINUITY, SOURCES, AND CONTRIBUTION OF VERTICAL MOTION FIELDS

In presenting the synoptic and mesoscale summaries, a clear correspondence has been established between the major convective complexes and regions of positive environmental vertical motion. Also, the analyses support the contention made in Chapter 4 (see Section 4.8), that the adiabatic technique used to compute vertical motion depends primarily on the environmental dynamics; that is, on motions going on outside of active convective cells.

However, the spatial and temporal continuity of the centres of ascent has not been clearly established, and neither has their source regions. In particular, did the vertical circulations leading to the AVE-V storm originate locally, or further southwest toward the major synoptic system? Either way, the associated convective events did not occur until several hours after the mesoscale circulations were established over each region! Since the mesoscale vertical motion patterns appear to be controlled, if not created, primarily by synoptic processes, they should also be predictable to a large degree. The clear implications for forecasting will be examined further in Chapter 6.

5.3.1 Spatial Continuity

The discussion has alluded to the close relation between surface convergence and the centres of ascent aloft. Hourly analyses of the surface pressure fields (Appendix IVa) did not demonstrate the pre-storm convergence conclusively. However, Sikdar and Fox (1983) have provided the necessary proof for the AVE-V case from surface divergence

computations using wind data (Figure 5.6f). An hourly analysis of surface divergence may help resolve the question of source regions for the mesoscale dynamics.

To this point, only isobaric analyses of the vertical motion fields have been discussed. Examination of the radar composites and satellite imagery reveals that the major storm centres were 200-300 km apart. The mesoscale upper air network, with a mean site spacing of 75-80 km, was therefore capable of resolving features separated by more than about 150 km. This made it possible to relate broad regions or ridge lines of ascent to concurrent convective complexes. Spatial continuity of the computed vertical motion fields was established, in the sense that the patterns could be related to patterns at other levels, to other dynamic features, and to the visible cloud formations.

However, temporal continuity in the isobaric centres of ascent was not demonstrated conclusively, since these centres could not be tracked continuously with any degree of certainty. This problem is not a result of the lack of temporal resolution in the data. Rather, it results from the mesoscale network domain limitations, and the lack of spatial resolution outside this domain, with resultant aliasing of data. In effect, the mesoscale network was too small to follow features of 200-300 km wavelength travelling at 30-50 km per hour for more than about 4-8 hours. In other words, many features were moving through the network within two analysis times.

Temporal continuity could equally not be established for the related secondary surface centres of low pressure (Appendix IVa), where one-hour temporal resolution was available, but here the spatial

resolution of data was only 130 km. The problem of temporal continuity will now be examined further using vertical cross-sections.

5.3.2 Temporal Continuity and Source Regions

For the AVE-V case, the main storm forms at the southwest extreme of the mesoscale network (Figure 5.9) with a southwest flow aloft (Figure 5.1). Since the pre-storm environment is one of the primary concerns of this study, the lack of spatial resolution of the soundings outside the mesoscale network is restrictive not only for establishing the temporal continuity of the centres of ascent, but also for determining possible source regions of these centres. The coarse synoptic resolution (370 km on the average) results in maxima and minima in the vertical velocity occurring at the synoptic stations (see Figures 5.6-5.16). Corrections to the locations of these errant centres of vertical motion can be applied by examining space and time cross-sections of the vertical motion.

The lines and sites chosen here for the space-height and time-height cross-sections are depicted in Figure 5.17. The north-south line, from Dodge City, Kansas (DDC) to Victoria, Texas (VCT), runs normal to the surface front (Appendix IVa). The time series of vertical cross-sections for this line appears in Figure 5.18. The next set of cross-sections (Figure 5.19) is more or less parallel to and north of the surface front throughout the AVE-V period, running west-east from Albuquerque, New Mexico (ABQ) to Monett, Missouri (UMN). A third series of cross-sections running from MAF to UMN, is presented for the first

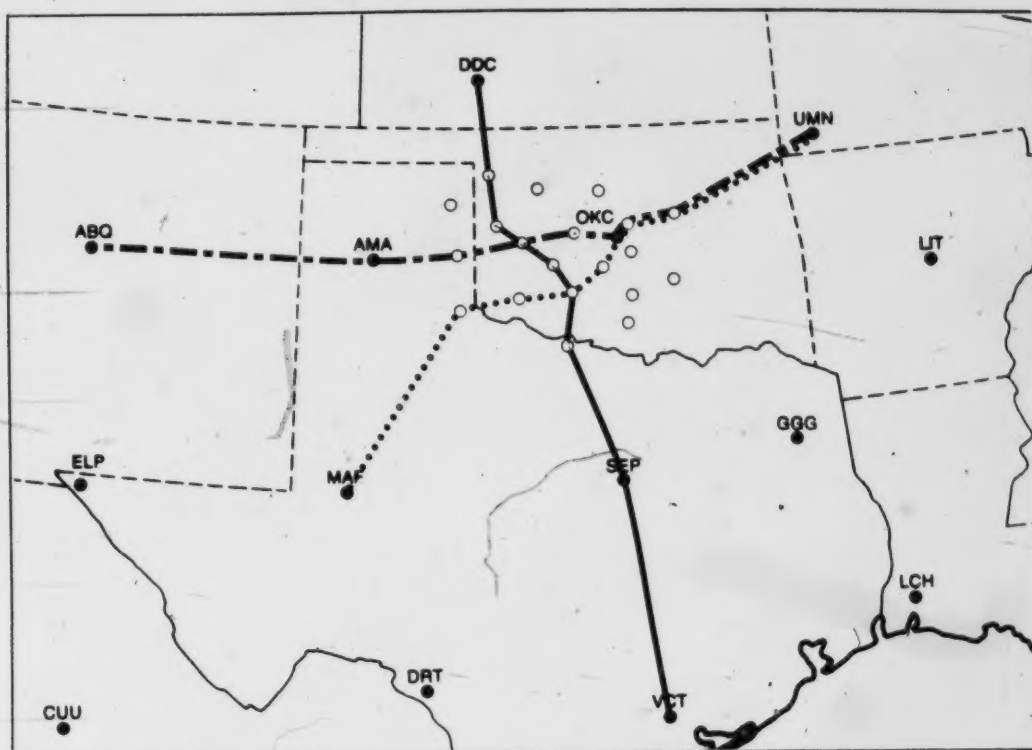


FIGURE 5.17: Vertical cross-section lines used for Figures 5.18 - 5.20.

half of the storm period (Figure 5.20). This cross-section parallels the southwest flow aloft as well as the general motion of the convective storms, but it also contains sites close to the initiation regions of the AVE-V storm complex (southwest Oklahoma) and the Midland, west Texas storm. The fourth series presented in Figure 5.23 is a set of time-height cross-sections of vertical motion for eight selected sites. These are particularly useful for narrowing down times when particular features in the vertical velocity field passed over a given site.

These space and time cross-sections will now be used to examine the temporal continuity of the vertical motion patterns. At the same time, an attempt will be made to determine the primary source(s) of some of the centres of ascent noted in Figures 5.6-5.16. A specific objective here is to determine whether the centres of ascent associated with the AVE-V complex are primarily:

- a) generated locally, solely as a result of convective processes and/or orographic lift;
- b) generated locally as a result of the front moving southward and lifting the warmer airmass;
- c) propagating sub-synoptic or mesoscale circulations, spawned upstream as a consequence of a downscale transfer of energy from the synoptic scale to the observed scale of motion.

If the relative importance of these processes can be resolved, one might then proceed to relate sources and sinks of synoptic scale energy to the severe storms.

Convective processes were earlier discounted as the major source for the observed centres of ascent for the AVE-V complex and other major storms of this day, since low-level convergence and low- to mid-level ascent was already present over each region 4-6 hours prior to first radar echoes (Figure 5.6). At the same time, these centres of ascent appeared to move from southwest to northeast. And while convective motions usually enhance the local environmental vertical motion field once a storm forms, the analyses of Section 5.2 suggest that convection is not a primary source for the mesoscale environmental ascent and surface convergence discussed here.

The cross-sections of Figure 5.18 (DDC to VCT) are normal to the surface front. These are first examined in order to investigate whether the front caused the observed centres of ascent and the associated storms to form. If such were the case, one would expect southward propagation of the ascent centres similar to the frontal motion. The main storm (AVE-V complex) was initiated just west of the cross-section line and passed over it during the period of peak intensity (202000 to 202200). The frontal position super-imposed on each cross-section of Figure 5.18, was identified using the surface analyses of Appendix IVa, strong horizontal temperature gradients at 850 and 700 mb, and significant wind shear lines.

While the surface front moved southward from Mountain View (MTV) to near Stephenville, Texas (SEP) between 201230 and 210330, the broad meridional patterns of vertical motion moved little during the period. That is, low-level ascent over the Gage - Wichita Falls (GAG-SPS) region at 201230, was evidently still over the same vicinity at 210330, except for changes in intensity as the storms moved eastward perpendicular to the line. Subsidence at higher levels over the GAG-MTV segment at the start, also remained at 210330. One therefore sees little in these patterns to suggest that the southward movement of the front resulted in the observed centres of ascent. After 210330, more significant synoptic scale changes appear to take over and finally change the patterns, with increasing ascent at higher levels.

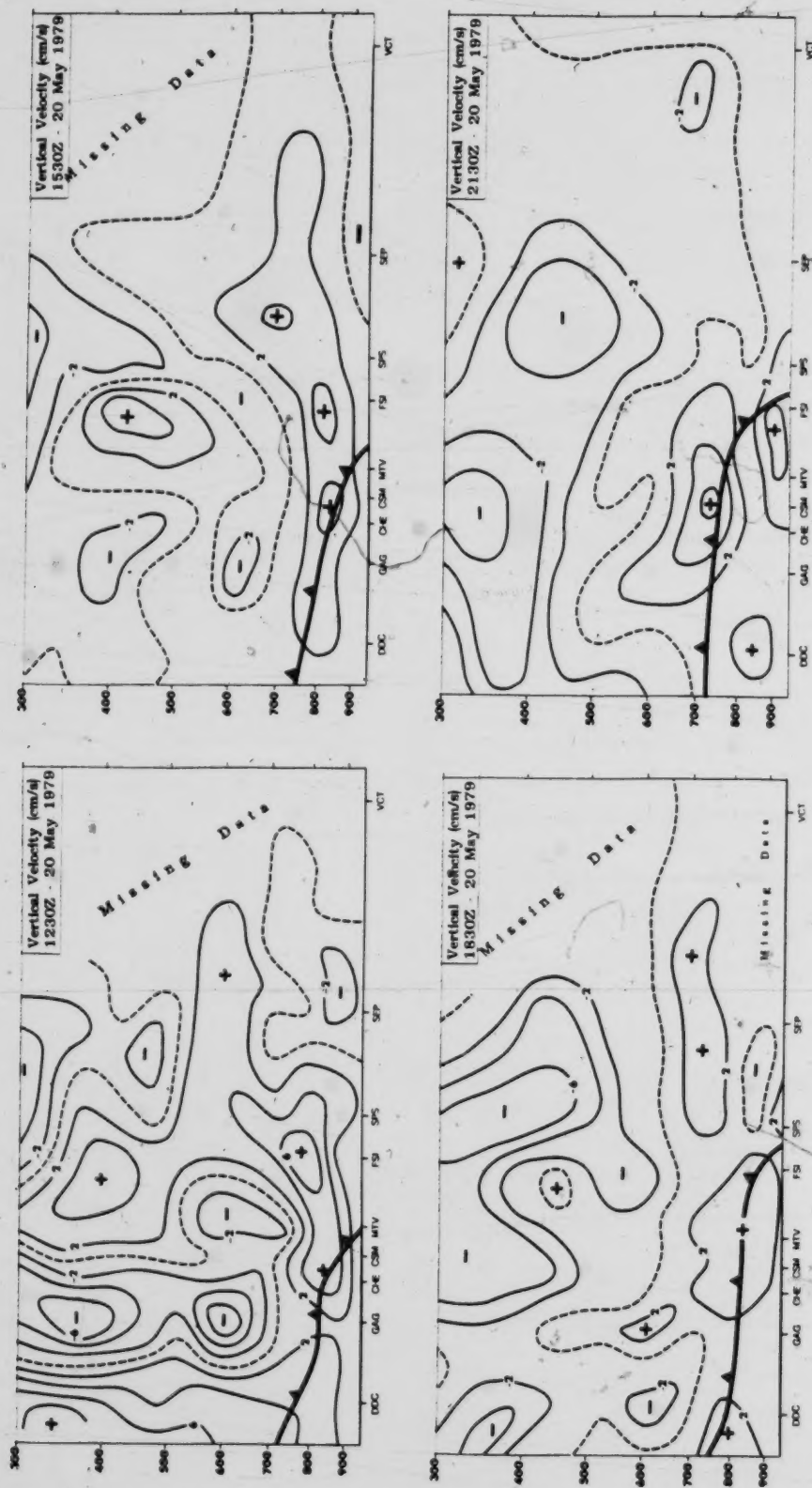


FIGURE 5.18: (caption next page).

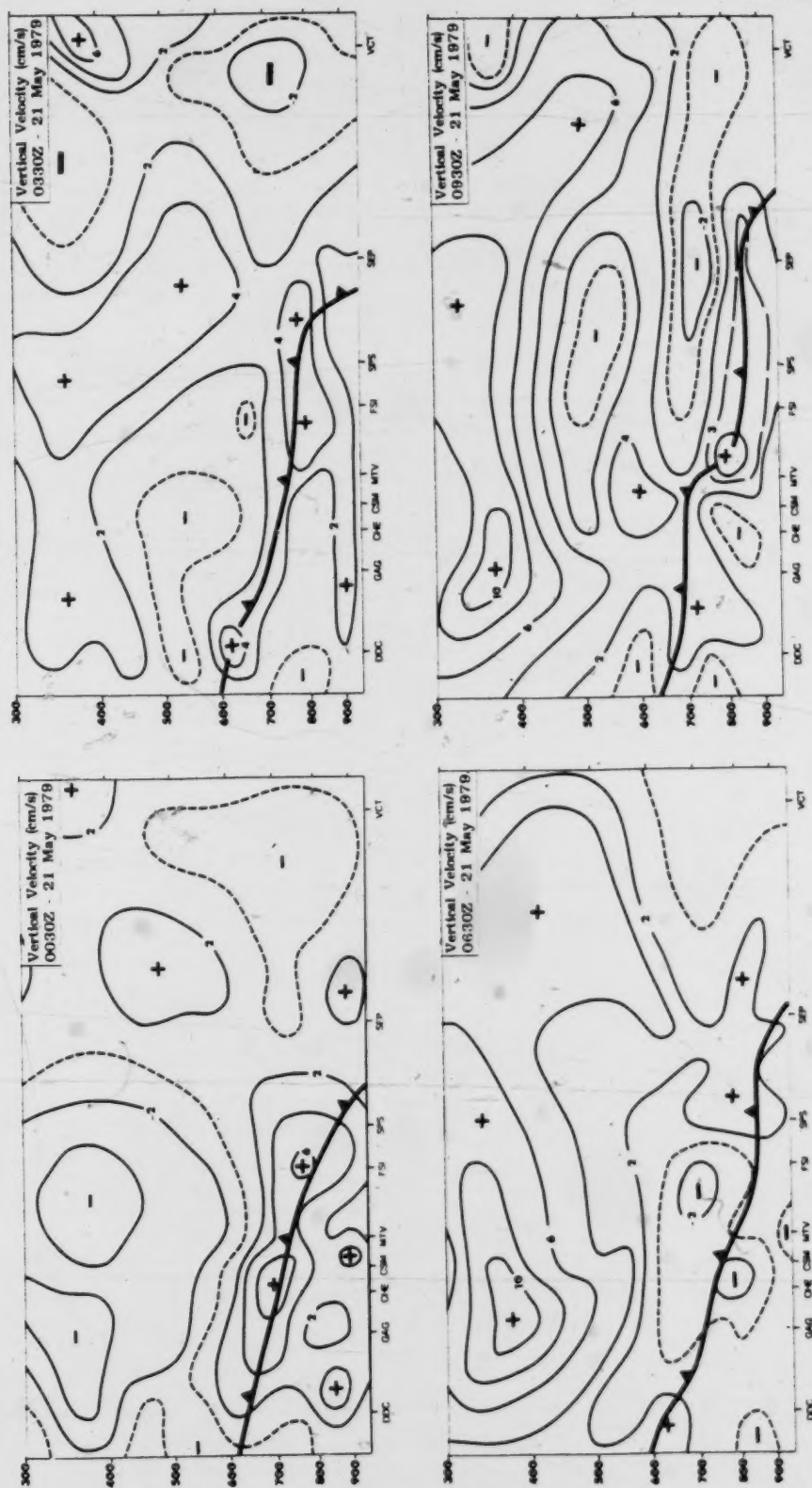


FIGURE 5.18: Time series of north-south (DDC - VCT) vertical cross-sections of vertical velocity, 1230-0930 UTC, 20-21 May, 1979. Broken lines indicate 0 cm.s-1, heavy solid lines the approximate leading edge of the cold front.

Some important general observations concerning these cross-sections can now be made. First, significant low-level ascent ($3-8 \text{ cm s}^{-1}$), concentrated below 800 mb, was present over regions where major convective complexes later formed. For the AVE-V storm, this ascent was present at least six hours before first radar echoes. This is evident in the space-height cross-sections of Figures 5.18 to 5.20, as well as in the relevant time-height cross-sections (Figure 5.21). This ascent also corresponds well with the regions of surface convergence computed by Sikdar and Fox (1983) (Figure 5.6f). Lewis et al. (1974) found low-level convergence 2-3 hours before radar echoes, but with a horizontal extent of only 50 km. The regions of ascent noted here are 150-300 km across. The difference in scale may be a function of data resolution, but the important point in both cases, is that most of the pre-storm atmosphere below 800 mb was undergoing weak to moderate ascent. This low-level ascent is believed to be critical to convective storm formation, since it initiates the gradual removal of the capping inversion, a key factor in the release of trapped moisture and subsequent convection.

Both spatial and temporal variations in the level of maximum ascent are conspicuous. The north-south cross-sections (Figure 5.18), for example, indicate a northward tilt with height of the centre of maximum ascent during the early to mature storm period (201830 - 210330). The time-height cross-sections (Figure 5.21), which are arranged in north-south pairs (AMA-MAF, SHM-CDS, CSM-SPS, and OKC-EMC), indicate that the northernmost sites had their main centres of ascent in the 750-700 mb layer, and the southern sites in the 850-800 mb layer. Over the mesoscale network, the time-height cross-sections also exhibit a

tendency for these low-level centres of ascent to rise once the storms formed, perhaps reflecting some influence of storm-scale ascent at or above cloud base level.

The cross-sections support the earlier qualitative comparisons of isobaric analyses of vertical motion with satellite and radar images (Figures 5.6 - 5.16). For instance, the ABQ-UMN cross-section for 202130 (Figure 5.19d) shows mid-level ascent corresponding to the storms across southern Oklahoma (Figure 5.11), with subsidence downwind over extreme eastern Oklahoma, a narrow break near the western Oklahoma border where there was also low-level subsidence, and ascent again where there were smaller severe storms over the eastern Panhandle and New Mexico.

Additional clues as to the origin and motion of the main centres of ascent (and subsequently of the convective storms) are also provided by the cross-sections. Two main centres of low- to mid-level ascent, corresponding spatially and temporally to the two main convective regions of the AVE-V complex, are evident on the ABQ-UMN and MAF-UMN cross-sections (Figures 5.19 and 5.20) from 201830 until 210330.

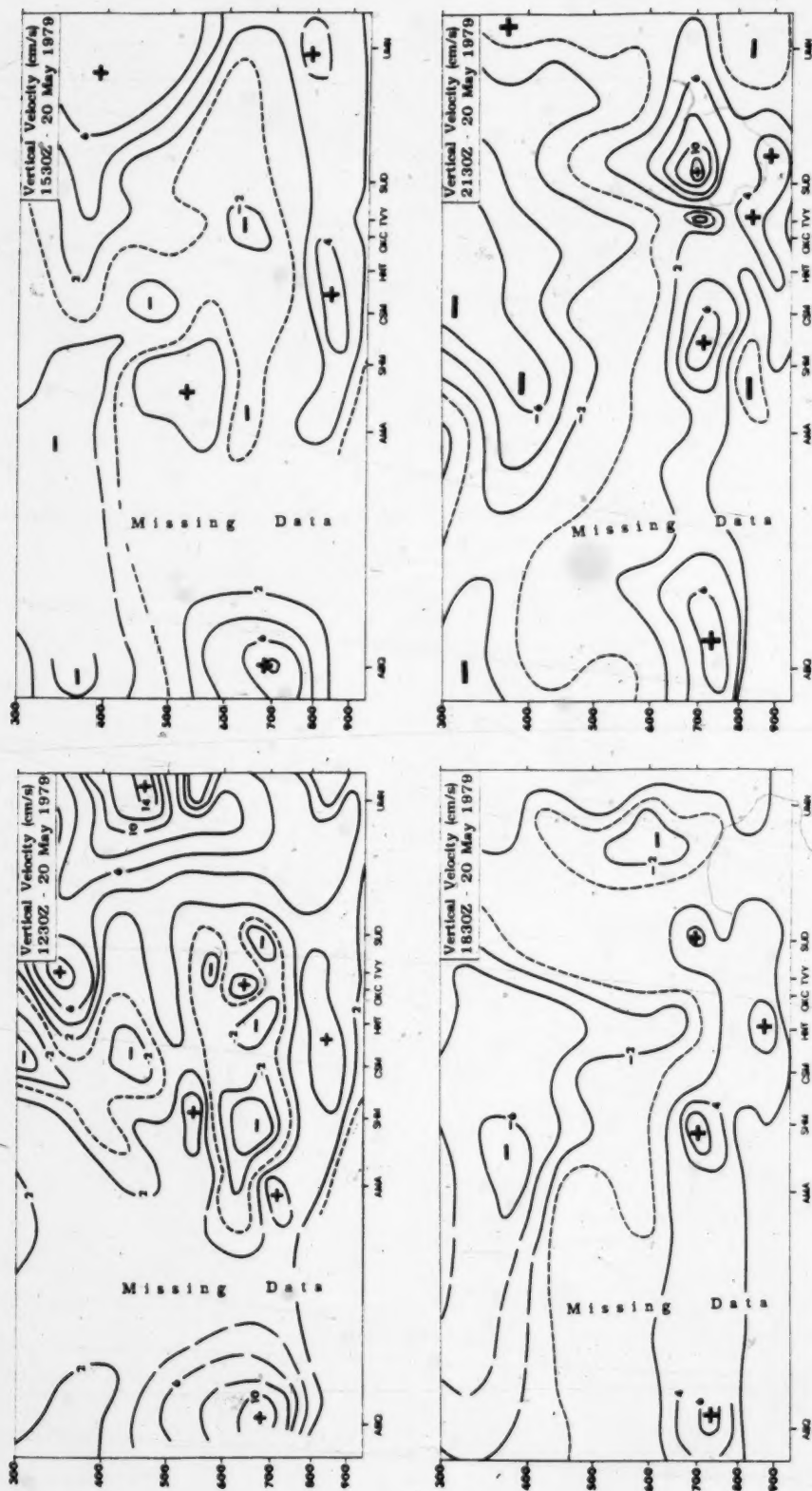


FIGURE 5.19: (caption next page).

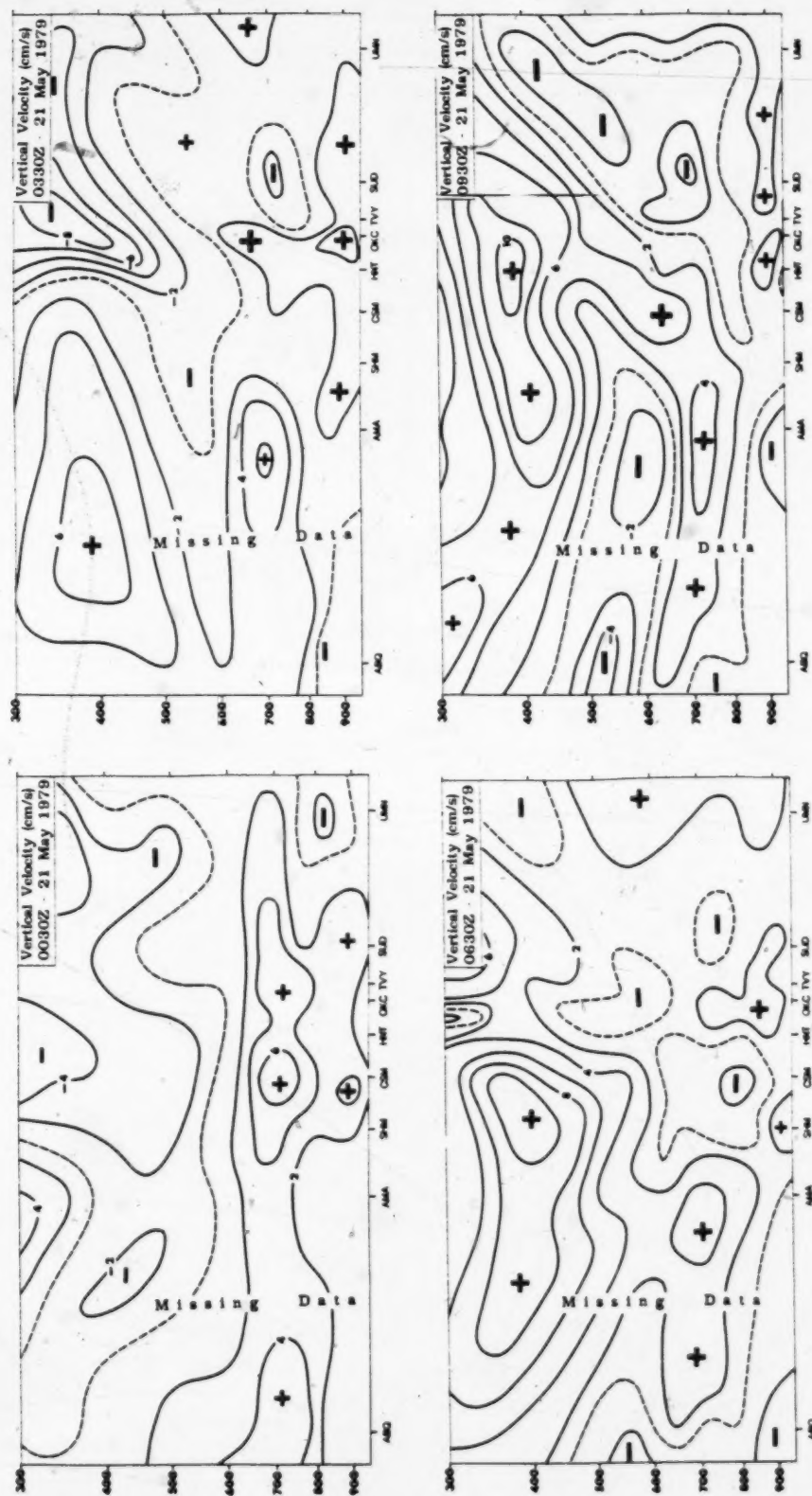


FIGURE 5.19: Time series of west-east (ABQ - UMN) vertical cross-section of vertical velocity, 1230-0930 UTC, 20-21 May, 1979.

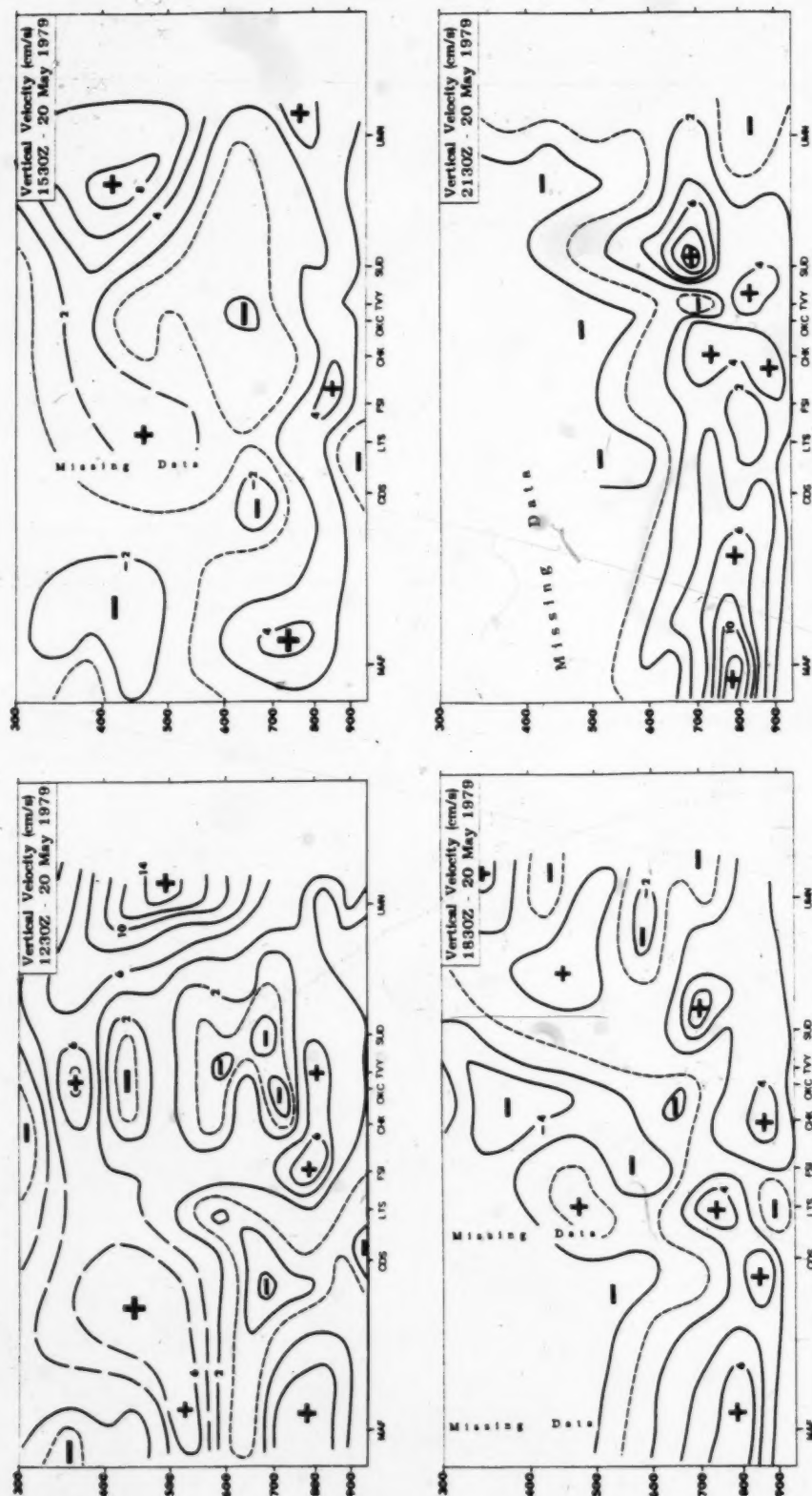


FIGURE 5.20: (caption next page).

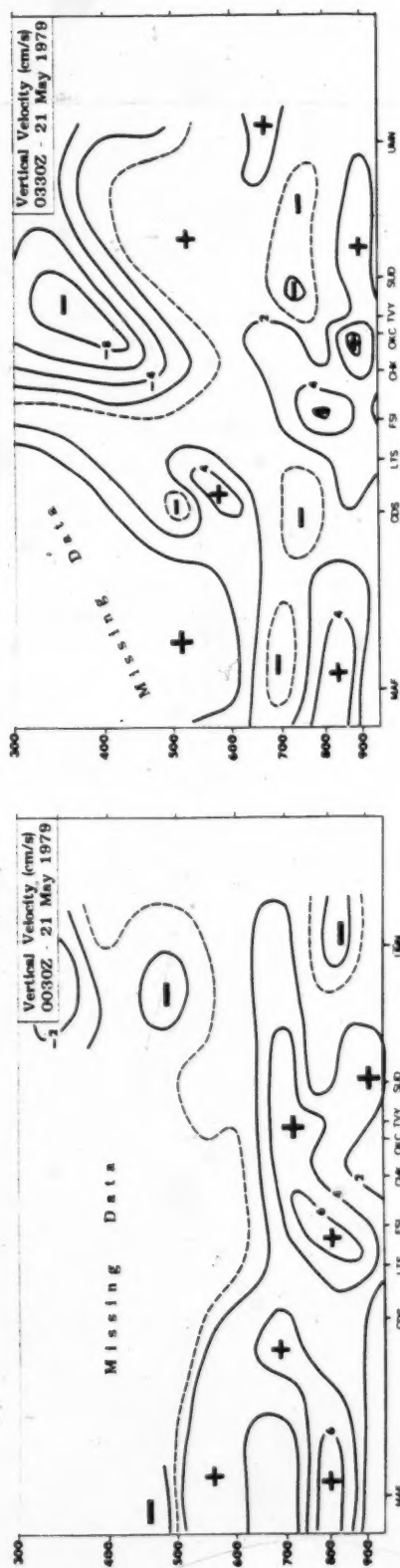


FIGURE 5.20: Time series of southwest-northeast (MAF - UMN) vertical cross-section of vertical velocity, 1230-0930 UTC, 20-21 May, 1979.

The time-height cross-sections (Figure 5.21) were used to determine the times of passage of significant centres of ascent. Combined with Figure 5.20 (MAF-UMN), these indicate that the eastern centre of ascent originated south and east of Wichita Falls (SPS) sometime before 201830 (time over Heaton - HEA), and appeared to move northeast at 30-35 km hr⁻¹. The more significant centre of ascent, indicated at 800 mb between CDS and LTS at 202130, originated west and south of CDS (southwest of Oklahoma) prior to 201830. Because of its limited history over the mesoscale network, one can only estimate a speed of 30-35 km hr⁻¹ towards the east. Of the two radiosonde sites west of the mesoscale network (AMA and MAF), MAF indicates the most significant centre of ascent prior to 201830 (Figure 5.21b). Using an estimated time of passage over MAF of 201300, a northeast direction of motion towards CDS gives a speed of 35-40 km hr⁻¹. Storm motions derived from radar observations were in the range of 30-50 km hr⁻¹ (July and Turner, 1981).

This reconstruction of the motion of this main centre of ascent is supported by the satellite imagery of mid-level clouds. A south-north oriented band of mid-level cloud west of MAF at 201300 (see satellite image, Figure 5.3 or 5.6), moved northeastward to the vicinity of CDS by 201830 (Figure 5.9), close to where the major AVE-V complex was initiated. This cloud band and its motion are particularly evident when viewed on an animated film loop of the GOES imagery.

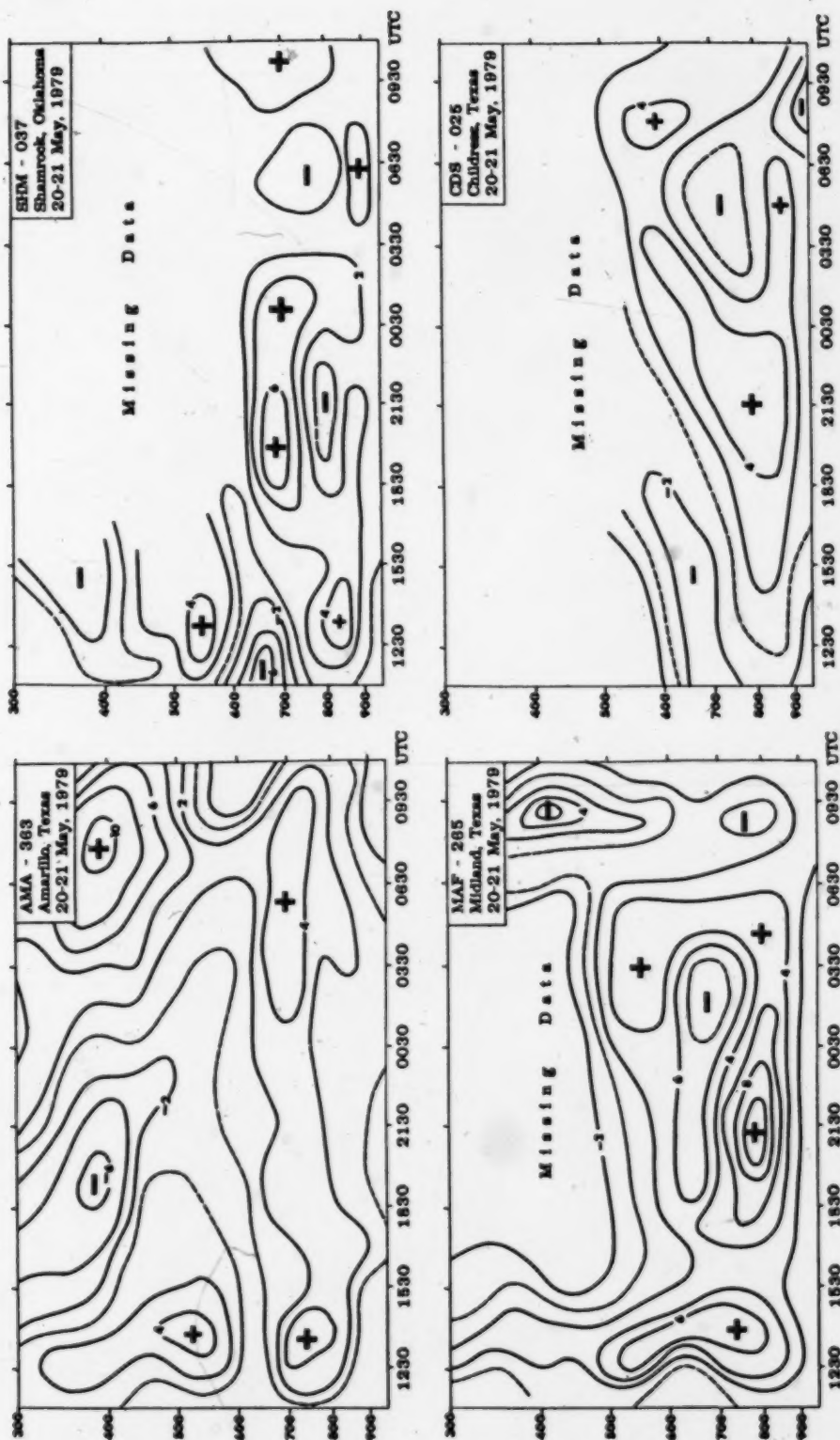


FIGURE 5.21: (caption next page).

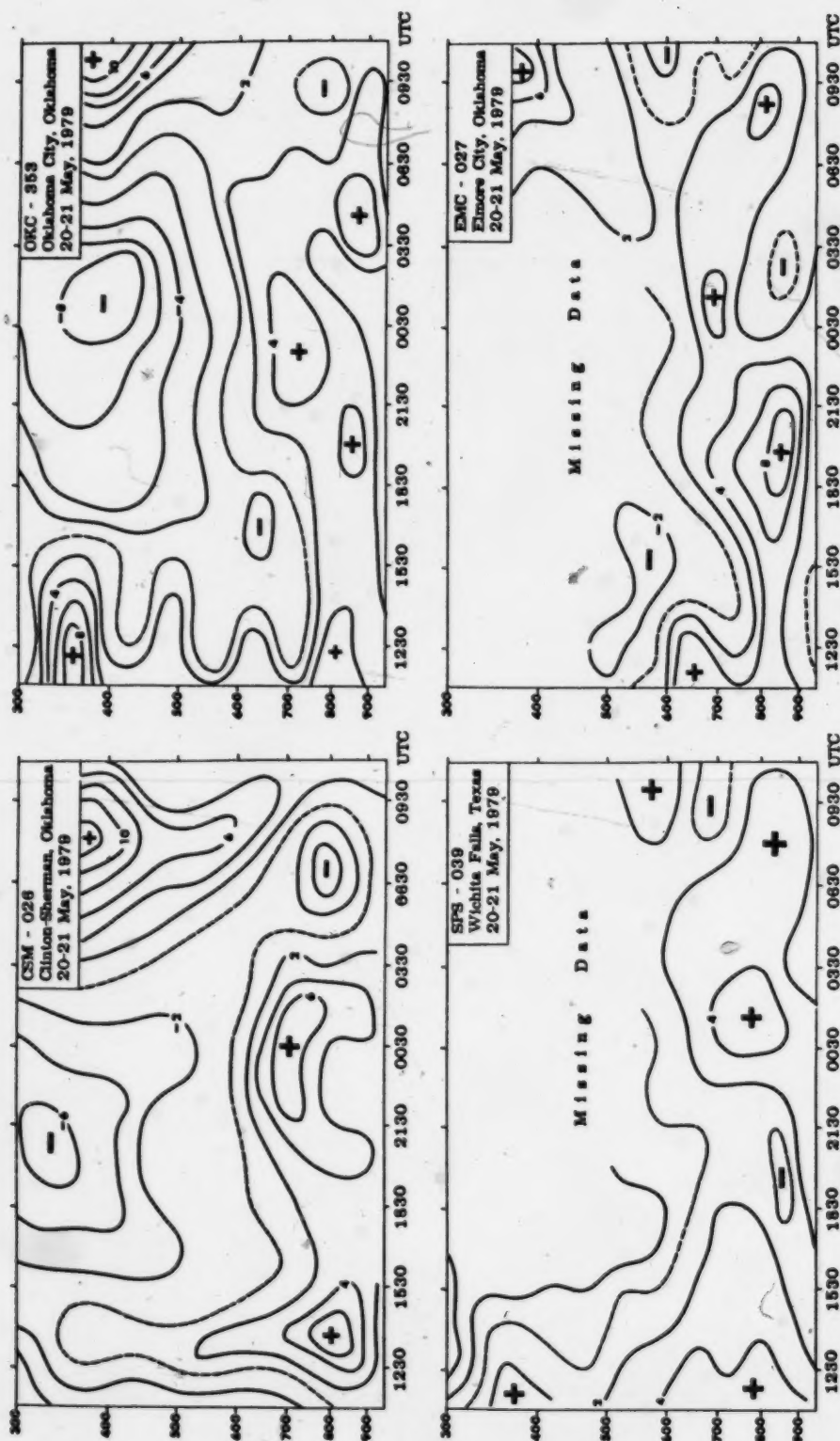


FIGURE 5.21: Time-height cross-sections of vertical velocity for selected sites, 1230-0930 UTC, 20-21 May, 1979.

5.3.3 The Contribution of Vertical Motion to Convective Storms

Taken together, the above clues form the following scenario. A region of low- to mid-level ascent with dimensions of 150-300 km, appeared to originate from the major synoptic system over northern Mexico, and was over southwest Texas (near MAF) at the start of the AVE-V data period. This feature subsequently moved northeast at $35\text{--}40\text{ km hr}^{-1}$, its advance providing adiabatic cooling of the lower atmosphere, particularly of the capping inversion layer. The energy contained in such regions of ascent might be looked upon as large-scale energy consumed by the storm. Its effective contribution to convective parcel energy should be proportional to the area of environmental cooling on an adiabatic chart.

The timing of these regions of ascent, together with the diurnal heating cycle and mesoscale pattern of boundary layer moisture, contributed significantly to the observed sub-synoptic scale (AVE-V) convective complex over southwest Oklahoma. Similar features aided the formation of other severe storms on this date, particularly the second AVE-V storm centre over south-central Oklahoma, and storms over the southern Panhandle and near Midland, west Texas. A more complete picture will emerge once the influences of the capping lid, wind fields, and moisture sources are discussed. Topographic vertical motion, although not computed directly in this study, is a component of the adiabatic vertical motion field used in this study. While its importance is unknown in this study, the migratory nature of the fields studied here, their relation to the synoptic scale, and the spatial scale of data used, lead one to believe that orographic ascent was not the important process in the initiation of the storms discussed.

5.4 CLOUD LATENT HEAT EFFECTS ON VERTICAL MOTION PATTERNS

It was anticipated in Section 4.4.4 that latent heat release from clouds could cause errors in the vertical velocity computations. Penetration of clouds by the radiosonde instrument would lead to under-estimates of ascent and over-estimates of subsidence when applying the simplified adiabatic technique adopted for this research. The vertical cross-sections just discussed raise this suspicion for the levels above 700 mb, especially during the mature storm period, when large areas of near overcast mid- and high-level clouds exist over and downwind of the severe storm regions. The exact cloud conditions at the locations and times of radiosonde releases are required in order to investigate this possibility further. To estimate cloud types, bases, thickness, and horizontal extent, one can utilize the regular surface observations, satellite imagery, radar reports, and temperature/dew point depressions ($T-T_d$) from soundings. None of these are accurate by themselves, since all are indirect methods of estimating the clouds. For example, visual surface observations of cloud base are not always reliable for clouds based above two kilometers. Satellite imagery should only be used to determine the horizontal extent of the clouds, once the types and elevations have been ascertained by other means. And, one cannot always rely on the inspection of moist layers of a sounding alone to infer cloud layers. However, all three of the above remote techniques are used to improve cloud estimates in the following section.

5.4.1 Determination of Cloud Types and Bases

Surface-based, satellite, and radar observations of the AVE-V clouds were earlier used to determine the synoptic scale cloud patterns and types, and summarized in Figure 5.4. To further refine these interpretations, Table 5.1 summarizes the values of $(T-T_d)$ used to determine cloudiness at the mandatory levels, a slight modification of those given by Rogers and Sherr (1966). Figure 5.22 provides an analysis of dew point depression $(T-T_d)$ at four mandatory levels (300, 500, 700, and 850 mb) for the three main synoptic times (1200, 0000, and 1200 UTC). The hatched regions then represent "cloudy conditions" (broken to overcast layer) based on the values of Table 5.1.

TABLE 5.1: Maximum dew point depression for cloudy conditions as a function of pressure¹².

Pressure	$(T-T_d)$ for given Pressure Levels					
	> 800	700	600	500	400	300 (mb)
(a) Rogers and Sherr (1966)	2	3	4	5	6	7
(b) this study	2	4	5	6	7	8

Based only on the criteria of Table 5.1, Figure 5.22 implies the presence of low (850 mb) and middle (700 mb) level cloudiness just northeast of the mesoscale network at 201100, with cirrus cloudiness (300 mb) possibly spreading into the southern half. The surface observations and satellite imagery suggest a similar picture. The 202300 patterns of Figure 5.22 suggest low cloudiness through northern

¹² These values should be applied with caution, as no statistical analysis was attempted for this relation.

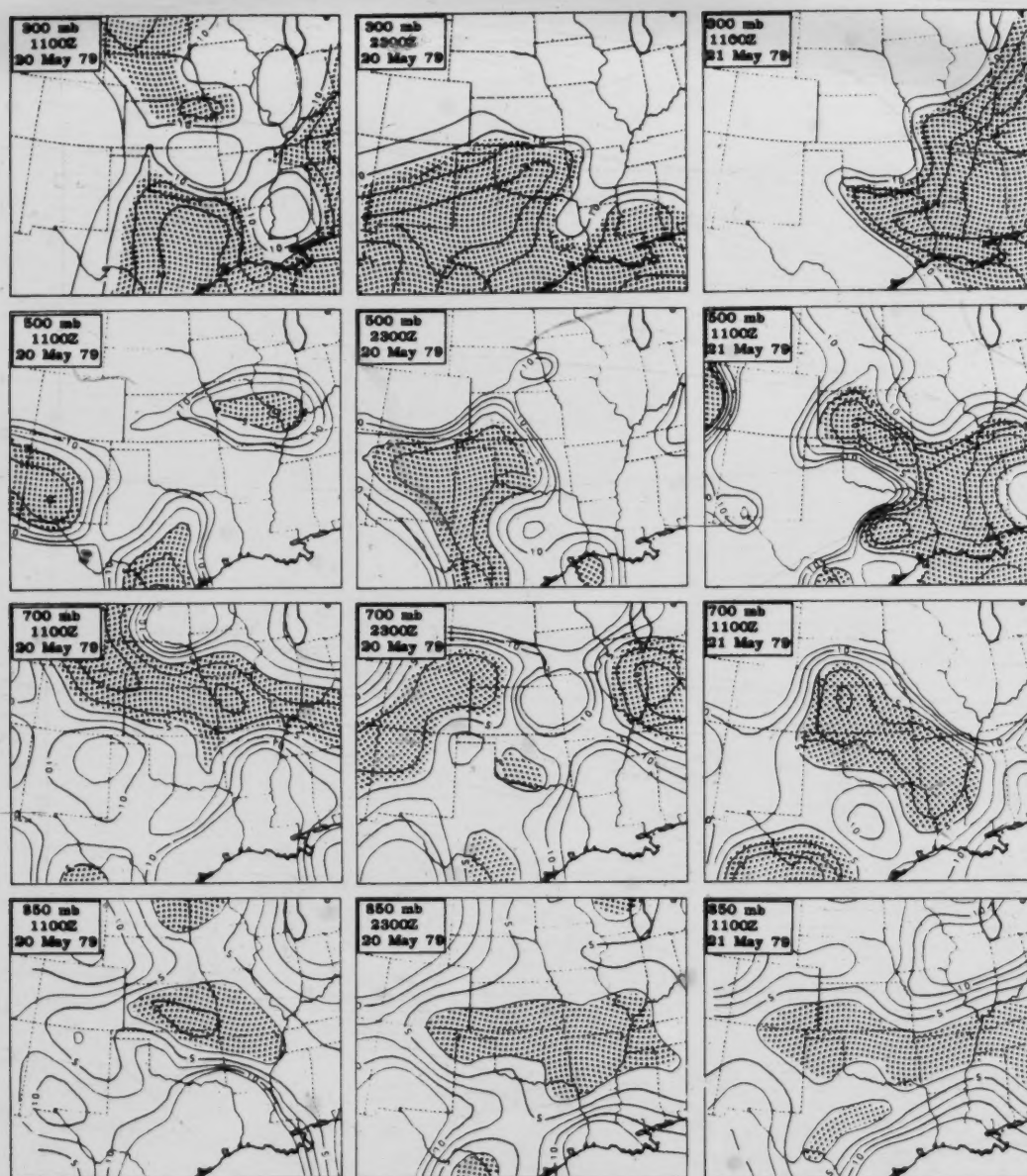


FIGURE 5.22: Dew point depression ($T - T_d$) for 300, 500, 700, and 850 mb, 1100-1100 UTC, 20-21 May, 1979. Contour interval is 2.5°C, stippled areas are potential "cloudy regions" based on the values in Table 5.1.

Oklahoma and the Texas Panhandle, 700 mb clouds over much of the AVE-V complex, and cloudy conditions at 500 and 300 mb throughout the western portions of the two states. By 211100, the patterns imply that most of Oklahoma had cloudy conditions up to middle levels, with the higher clouds having moved off to the northeast of the mesoscale network. Since this cloudiness appears to place the adiabatic technique of computing vertical motion in some jeopardy in this case, the problem requires further investigation.

5.4.2 Validity of the Computations

The time-height cross-sections of vertical velocity for both Amarillo (AMA - Figure 5.21a) and Clinton-Sherman (CSM - Figure 5.21c) exhibit a significant deep layer of subsidence above 600 mb for at least a portion of the mature storm period. Figure 5.23 shows soundings for these two locations for the period of significant subsidence (201700 and 202300), along with the temperature trace (dotted lines) for the subsequent sounding three hours later in each case (202000 and 210200). The CSM soundings are representative of the severe convection of the AVE-V complex. The AMA soundings are representative of the moderate convection which occurred west of the AVE-V complex.

For the Amarillo cross-section and soundings (Figures 5.21a and 5.23), there is little evidence to suggest that latent heat was erroneously interpreted as subsidence in the vertical cross-section profiles. The subsidence above 600 mb between 201700 and 210000 is evident in the satellite imagery, and it seems to have been a factor in preventing the AVE-V storms from forming northwest of Amarillo over the

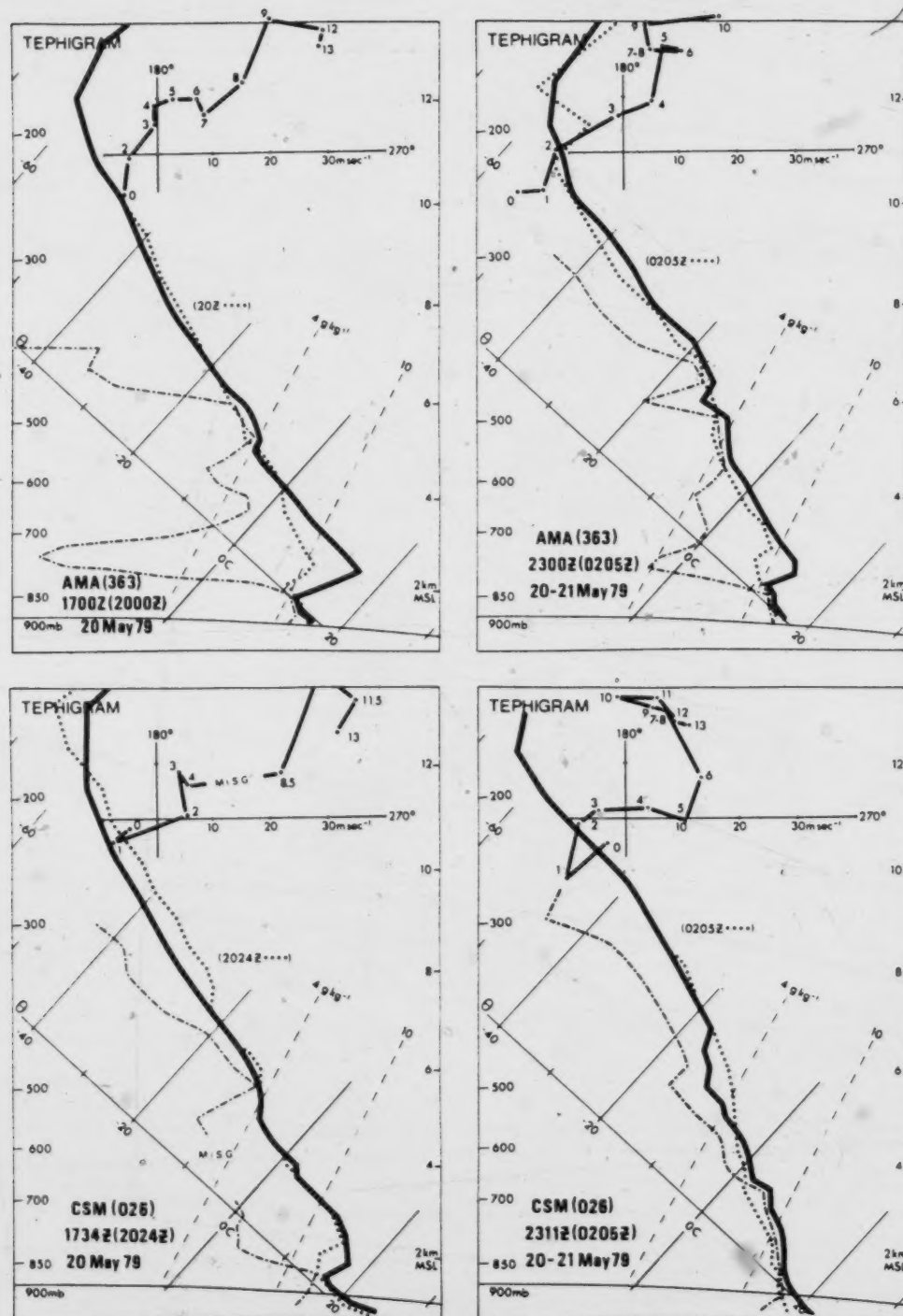


FIGURE 5.23: Soundings for Amarillo, Texas (AMA), and Clinton-Sherman, Oklahoma (CSM), 20-21 May, 1979. The solid lines represent temperature and winds (hodograph with altitudes plotted in km). The broken lines are dew point temperatures. The dotted lines are temperature soundings three hours later (actual times in parentheses).

Panhandle region during this time period. By the time convective storms did form or move over the northern Panhandle after 210000, a deep layer of ascent was already present.

On the other hand, the Clinton-Sherman sounding (Figure 5.23) and most of the other soundings of the mesoscale network, suggest some warming from latent heat release above 700 mb. Still, the subsidence over CSM indicated in Figure 5.21c commenced before 201530, several hours before significant upper cloud layers were present (after 201800), and more than four hours before radar echoes of convection. The satellite imagery of Figures 5.7-5.9 show a very definite subsidence signature (clear air) over the region as well. Since the SESAME soundings did not penetrate active convective cells (Sienkiewicz et al., 1981), one may infer that the subsidence layer above 600 mb did, in fact, exist. This subsidence may have been enhanced above 500 mb on the downwind side of the AVE-V complex by compensating storm scale subsidence after 201830. Also, because there appear to have been significant layers of cloud present over CSM from 900 to 300 mb during the severe storm period (201830 to 210300), it is reasonable to assume that the values of ascent below 600 mb were somewhat underestimated¹³, while the subsidence aloft was over-estimated, though to a lesser extent. However, the mesoscale environmental values of vertical velocity appear to be of the right sign, and there is no reason to believe that they are any less accurate than the uncertainties of Table 4.3 indicate.

¹³ For example, Fuelberg and associates (pers. comm.) computed 700 mb kinematic values of $10-12 \text{ cm s}^{-1}$ near CSM at 202130 (average of 202000 and 202300 values), nearly double those in Figure 5.21c; however, these include the area-averaged storm-scale motions.

The subsidence over CSM which spread downward to the boundary layer once the storms had moved eastward and dissipated (after 210330), is a real feature, supported by the prevailing clearing weather conditions, and it is reflected in the regional analysis of vertical velocity of Figure 5.15.

5.5 TIME- AND AREA-AVERAGED VERTICAL VELOCITY FIELDS

In Chapter 4, some significant differences were noted between the isobaric analyses of adiabatic vertical velocity as used in this study, and kinematic analyses as reported by others. In this section, we encounter even greater differences in vertical profiles of vertical motion between the two techniques. This necessitates further validation of the adiabatic technique.

The most notable features of the adiabatic computations of vertical velocity for AVE-V were low-level ascent, especially during the pre-storm period, and apparent subsidence above 600 mb, which lasted well into the mature storm period over the mesoscale network. In order to examine the persistence of such features, Figure 5.24 provides a time series cross-section of vertical velocity averaged over the entire AVE-V mesoscale network from 201230 to 210930. There are three major features on this cross-section.

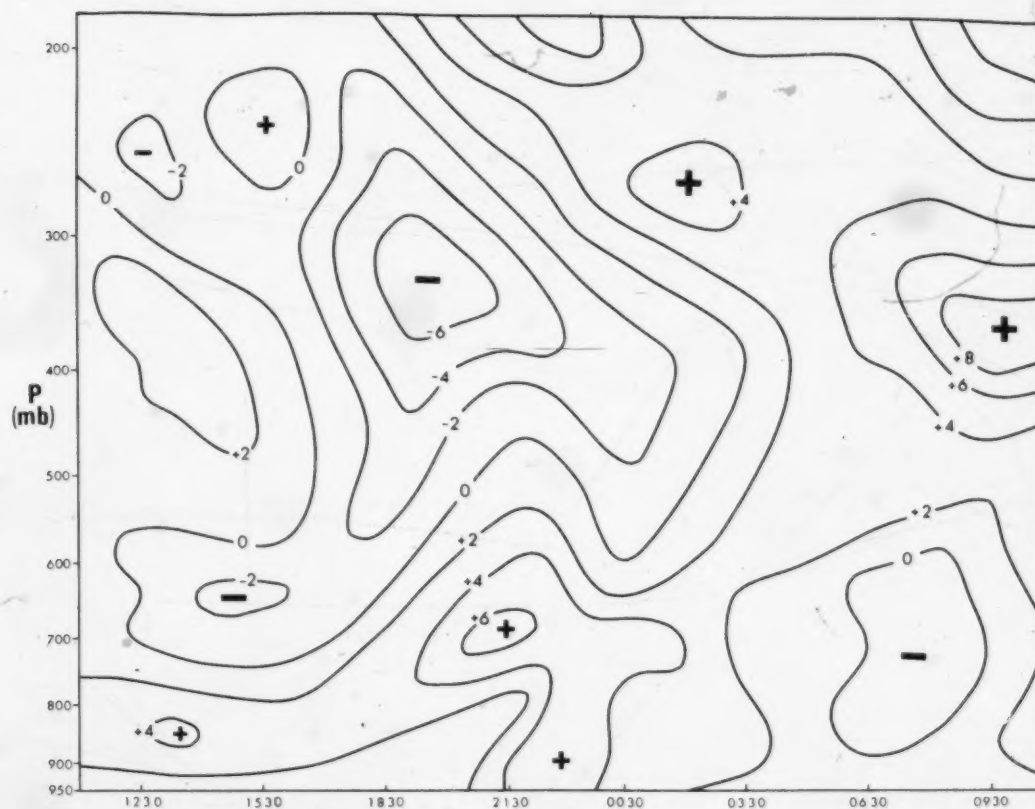


FIGURE 5.24: Adiabatic computations: time-height cross-section of area-averaged vertical velocity (cm s^{-1}) over the AVE-V mesoscale network, 20-21 May, 1979.

5.5.1 Pre-storm Ascent at Low Levels

First, low-level ascent, even when averaged over the whole mesoscale network, was a prominent feature of the pre-storm period (201230 and 201530), peaking near 850 mb. The peak rose above the 800 mb level by the time of storm initiation (201830), and to 700 mb or higher during the mature storm period (202130 - 210030). Ascent remained a dominant low-level feature during AVE-V until the decay phase (after 210500), when the storm was moving away from the mesoscale network as well as decaying, and weak subsidence had set in at lower

levels. The level of maximum ascent rose into the upper troposphere (above 400 mb) during this period. This latter feature appeared to result from the synoptic vortex approaching from the southwest, rather than from the dying convective complex off to the east. The upper vorticity pattern shown moving towards Oklahoma in Figure 5.5, while surface pressures were rising, would support this pattern.

5.5.2 Storm Period Subsidence at Middle Levels

The second feature of the Figure 5.24 time series is the persistence of subsidence above 700 mb. This appears weak at first, with a small peak at 650 mb at 201530, just prior to storm initiation. The intensity of subsidence then appears to increase, while its peak level rises above 400 mb during the early and mature storm stages (201830 - 210030). This subsidence feature, especially during the mature storm stage, appears to contradict kinematic studies which consistently show strong ascent near 400 mb during mature storm periods. Examples include Fankhauser (1969), Ogura and Chen (1977), and Fuelberg and Printy (1983). Since the latter study dealt specifically with this same AVE-V case, the differences appear to be a function of either the data base used (winds versus temperatures), or the method of computation. The question remains whether the adiabatic computations are in error because of latent heat release, and if so, is it significant enough to replace ascent by fictitious subsidence? However, it should be reiterated here that the kinematic method includes an area-averaged component of storm-scale motion, while the adiabatic method should only be sampling the environmental contribution to the

total. This rather drastic difference in vertical motion patterns between the two techniques can now be reexamined, using a kinematic time-series cross-section comparison with Figure 5.24.

5.5.3 Conflict with Kinematic Computations

A kinematic time series cross-section for the same AVE-V case is reproduced in Figure 5.25a (from Fuelberg and Printy, 1983), with values of vertical velocity in b s^{-1} . In the kinematic computations, one notes the strong influence of storm-scale motion during the mature phase. Apart from the opposite signs of the vertical motion fields at upper levels during the mature storm period, other features of the two (adiabatic and kinematic) vertical cross-sections are actually quite similar. This includes the pre-storm mid-level subsidence and weak low-level ascent, and the weak low-level subsidence during the late storm decaying stage.

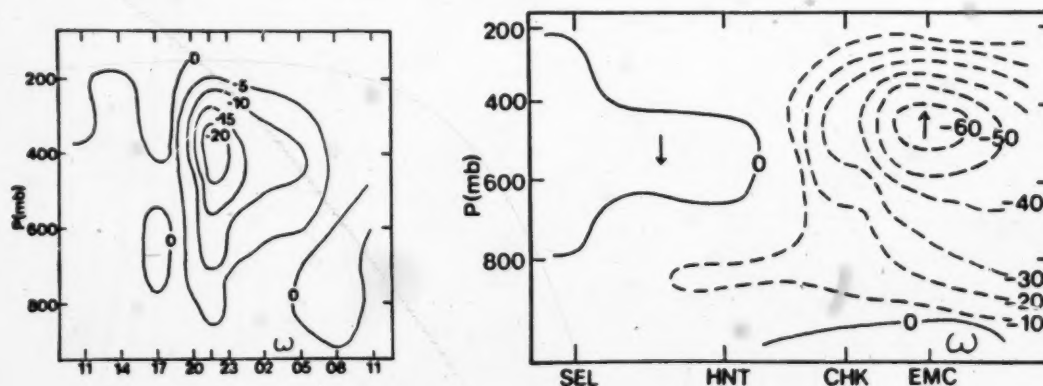


FIGURE 5.25: Kinematic computations: (a) time-height cross-section of area-averaged vertical velocity (b s^{-1}) over the AVE-V mesoscale network, 20-21 May, 1979; (b) space-height (SEL-EMC) cross-section of vertical velocity; 2130 UTC, 20 May, 1979 (reproduced from Fuelberg and Printy, 1983).

The accompanying space-height cross-section of Figure 5.25b (also from Fuelberg and Printy, 1983) shows that the kinematic ascent values of $20\text{--}60 \text{ m s}^{-1}$ near 400 mb extend over a region of at least 100 km, far larger than any individual storm cell. If one were to compact this rate of motion into a smaller region of, say, 20 km wide, this rate of ascent then translates to the order of $5\text{--}15 \text{ m s}^{-1}$. These values are more representative of ascent in the inflow regions of storm cells. Fankhauser et al. (1985), for example, using a combination of doppler and aircraft measures, computed vertical motions in this range of values over small areas of 5-10 km across in the storm inflow region of an active Montana storm cell. Values outside the storm inflow regions were less than the sensitivity of measurements ($< 1 \text{ m s}^{-1}$) would allow. On the other hand, cloud photography, parcel theory, and aircraft measurements, show that updrafts within severe thunderstorms, and ~~down~~drafts in precipitation shafts, are typically of the order of $10\text{--}60 \text{ m s}^{-1}$, and are usually confined to a small region within the storm. These latter scales of vertical motion, $5\text{--}15 \text{ m s}^{-1}$ in inflow regions, and $10\text{--}60 \text{ m s}^{-1}$ in thunderstorm updrafts, are one and two orders of magnitude greater than the kinematic computations of Figure 5.25.

Furthermore, cloud models show that the storm's return flow (subsidence) takes place primarily in the environment just outside the storm cell (Ogura and Cho, 1973). In other words, the storm-scale circulation appears to be confined to the storm and its immediate environment. This circulation alone, even though there are often several thunderstorms comprising one convective complex, should not significantly affect the environmental thermodynamics between storm cells, nor therefore, the adiabatic computations of vertical motion. At

the same time, however, the storm cells and the greater convective complex can and do affect the environmental winds over a much larger area, and therefore kinematic computations of vertical motion as well.

This argument supports the contention put forth in Chapter 4, that kinematic computations of vertical motion in the vicinity of convective storms cannot be considered accurate interpretations of either the environmental or storm-scale motions. Rather, they are area-averaged values of the total field, which are highly dependent on the spatial resolution of the data employed.

5.5.4 Vertical Velocity Features and Synoptic Processes

The third important feature of Figure 5.24 is the upward shift with time of the line of maximum subsidence (up to 202130), and of maximum ascent thereafter. The resulting tilt in the patterns is similar to what one should expect in the vertical profile of environmental vertical velocity, without the influence of convective storms. Initially, the layer of mid-level subsidence, which may result from local orographic flow or a shortwave ridge, can help create a capping lid effect. Later, low-level ascent commences as a shortwave trough approaches, and the ascent becomes an important feature leading to storm formation, since it results in surface convergence and the organization of low-level moisture, as well as the removal of any capping lid.

To further the argument on the validity of the adiabatic computations, it is helpful to review the synoptic pattern in yet another way. Figure 5.26a is the difference between the 500 mb

vorticity analyses of Figure 5.5; that is, the 12-hour vorticity change between 201200 and 210000. The regions of vorticity increase are reasonable approximations to the 500 mb region of synoptic scale environmental ascent at the mean time of 201800. By comparison with the time series cross-sections of vertical motion, the vorticity decrease across most of Oklahoma (Figure 5.26a) supports the 500 mb area-average subsidence of Figure 5.24. Similarly, the vorticity increase east of Oklahoma is reflected in the adiabatic cross-section with mid-level ascent prior to 201530. The features of Figure 5.26a continued a northeast motion, so that the approaching vorticity increases over New Mexico and west Texas should have strongly influenced the mesoscale network after 210330, 12 hours later. The adiabatic cross-section does indeed reveal strong ascent above 600 mb after 210330, once again supporting the validity of the adiabatic computations in this case.

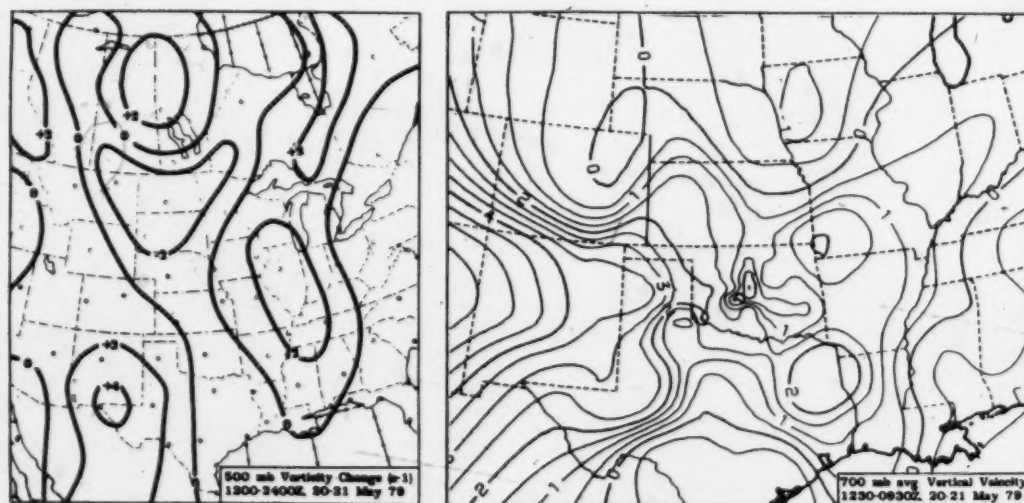


FIGURE 5.26: (a) 500 mb vorticity change (1200-0000 UTC), and (b) 700 mb 24-hour average (1230-0930 UTC) vertical velocity (cm s^{-1}), 20-21 May, 1979.

As a final comparison with the synoptic patterns of Figure 5.5 and 5.26a, time-averaged analyses of the isobaric vertical motion fields were investigated. One example of these is provided in Figure 5.26b, depicting the 24-hour (201230 - 210930) time-averaged 700 mb analysis, with a median time of 212300. The qualitative comparison with the vorticity change pattern is quite favourable indeed, except for slight differences in position between vorticity increase and ascent, and vorticity decrease and subsidence. This means that, in spite of computations in the vicinity of severe convective storms, the adiabatic method provides reasonable estimates of the environmental vertical motion field at the synoptic scale. In applying this method at the mesoscale, contamination by storm scale motion should be limited to the factors and uncertainties provided in Section 4.5.4.

5.6 SUMMARY: THE IMPORTANCE OF VERTICAL VELOCITY COMPUTATIONS

While trying to validate the adiabatic technique for environmental computations of vertical motion, doubts have been raised as to the interpretation of results of the kinematic method in the severe storm vicinity. Kinematic computations are significantly influenced by storm scale vertical motion, or at least by the effect of storms blocking or deflecting the wind flow in the storm region. This by itself does not invalidate the kinematic method, but it does place restrictions on the way that kinematic computations can be interpreted, and points out a need to obtain more accurate estimates of environmental vertical motion around severe storms. We seem to be capable of computing environmental vertical velocity in the absence of convective storms with reasonable

accuracy, using any of the standard methods. Values of storm scale motion within storm cells (updrafts and downdrafts) are also fairly well documented. However, short of using the adiabatic method, it is difficult to obtain reliable environmental measures of vertical velocity in the immediate vicinity of major convective complexes.

The energetics of interactions involving mesoscale storms and synoptic scale processes are poorly understood, and are therefore one of the most important areas requiring study. To study these interactions, it is necessary to obtain accurate information on the three-dimensional motion field. It is therefore vital to obtain more reliable vertical motion measurements or computations in future mesoscale experiments.

"As I began to learn meteorology, I found it necessary to unlearn some mathematics...."

"...Of course, even the purest of mathematicians use intuition to obtain preliminary estimates, but in meteorology these estimates are often the final product."

E.N. Lorenz, 1984:

speaking on "Irregularity: a fundamental property of the atmosphere".

CHAPTER 6

AVE-V SYNOPTIC TO MESOSCALE COUPLING

6.1 MESOSCALE WIND FIELDS

The initiation of severe convective storms is favoured over regions where a low-level wind maximum intersects either the left exit region of a cyclonically-curved upper jet, or the right entrance of an anticyclonically-curved jet (Section 1.3.3). In the ideal case, maximum upper and lower wind speeds of 40-60 and 20-30 m s⁻¹ respectively occur in the jet streak cores near the severe storm region (Uccellini and Johnson, 1979). Profiles of winds veering with height, similar to Figure 6.1 (from Chisholm and Renick, 1972) result, and these dynamics lead to differential heat and moisture advection. The upper jet aids convective development if it is situated so as to advect colder, dry air into the mid-to upper troposphere over the potential storm region, while transporting sensible heat downstream away from the region. These combined factors enhance upper level divergence leading to ascent below. Often, a low-level jet is coupled dynamically to the upper jet through an indirect circulation (Uccellini and Johnson, 1979). This helps to initiate and maintain convective storms by transporting heat and

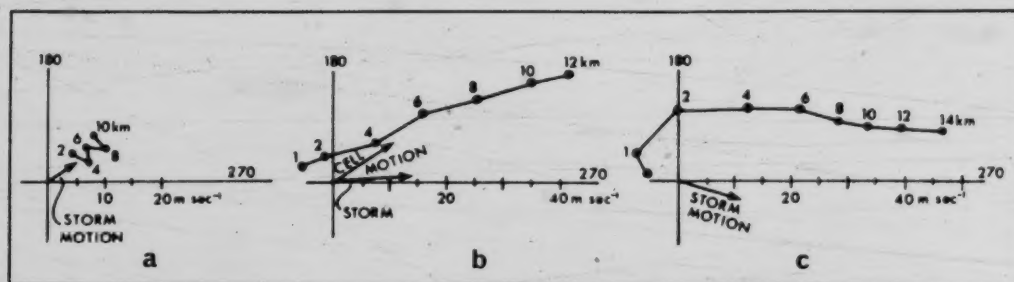


FIGURE 6.1: Typical hodograph wind profiles for Alberta (a) single cell, (b) multicell, and (c) supercell thunderstorms (after Chisholm and Renick, 1972).

moisture toward the region. Or, stated in another way, the low-level jet causes a cross-isobaric flow toward lower pressures, which leads to a conversion of APE to kinetic energy according to Equation 2.8.

In this section, we attempt to link the AVE-V mesoscale vertical circulations, or regions of ascent and subsidence as described in Section 5.2, to upper and lower jets, and thus corroborate these concepts of jet coupling. Figure 6.2 depicts analyses of the 300 and 850 mb height, temperature, and wind fields, using the additional spatial resolution of the AVE-V data period during the main synoptic hours. The 300 mb jet streams and 850 mb wind streaks were drawn with the aid of detailed wind isotachs (shown in Appendix IV-b), reported wind directions, the height contours (geostrophic flow), and cirrus streaks from the GOES visible imagery.

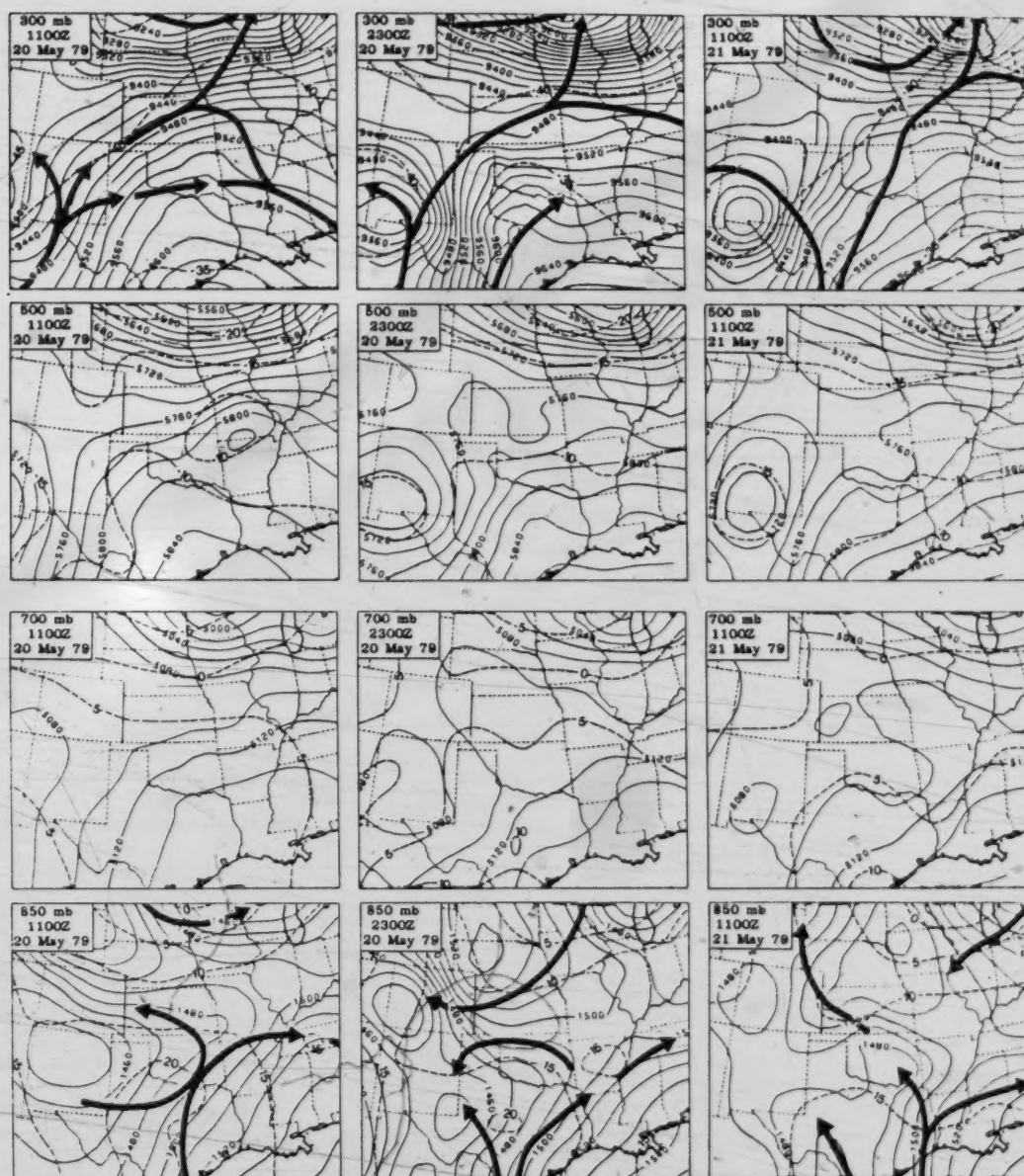


FIGURE 6.2: Mesoscale upper air analyses of height contours, isotherms, and upper and lower jet streaks at 850, 700, 500, and 300 mb, for 201200 - 211200, May, 1979.

6.1.1 Examples of Jet Streak Coupling

At 201100, the main 300 mb jet stream¹⁴ (UJ1 in Figure 6.2), had anticyclonic curvature with wind speeds in excess of 30 m s^{-1} over southern Oklahoma. A second anticyclonic jet streak (UJ2, northwest of UJ1) was breaking northeast towards the Great Lakes at this time. The northwest edge of the bands of cirrus cloud streaks appearing on the visible satellite imagery (Figures 5.3 and 5.6-5.13), also demarcate these jets. The low-level (850 mb) wind field has a broad maximum (Appendix IV-c), and one can resolve three weak jet streaks, LJ1, LJ2, and LJ3 as shown, with wind speeds of up to 20 m s^{-1} .

Three upper/lower jet streak intersection regions can be identified at this time: LJ2 and UJ1 over south-central Oklahoma, LJ1 and UJ2a over extreme southern Missouri, and LJ2 and UJ2 over southwest Kansas. Following the Beebe and Bates conceptual model, since upper jet UJ1 has anticyclonic curvature, the first of these intersection regions favours convective formation under the right entrance region of UJ1 over northwest Texas (Midland, to southwest Oklahoma). Similar arguments suggest convective formations over southwest Missouri and southwest Kansas. In addition, the convergence (Figure 5.6f) caused by the merger of LJ1 and LJ2 over south-central Oklahoma favours a fourth region. All four regions can be related to mid-level (700 mb) centres of ascent (Figure 5.6), as well as to convective storms either already in progress (southwest Missouri and southwest Kansas), or about to form. However, one would be reluctant to forecast such mesoscale storms in an operational setting, based only on the rather weak low-level jets in

¹⁴ The jets discussed briefly in Section 5.1.1 were based on synoptic analyses of the 250 mb level.

this case (see isotachs, Appendix IV-b).

A major reorganization of these jet structures occurred by 202300, perhaps influenced by the convective storms, 6-8 hours old by this time. The southern jet (UJ1) had weakened and slipped further south, giving lighter winds over southern Oklahoma (Figure 6.2). UJ2 was now the main jet, but it had diminished in intensity over and upstream (southwest) from the major convective storm region, while the 300 mb winds had generally increased by up to 15 m s^{-1} north of Oklahoma, downstream from the AVE-V storm. These changes in wind intensity will be shortly scrutinized further.

Two distinct southern low-level jet streaks, LJ1 and LJ4 can be identified at this time. Their locations are based on the surface pressure and 850 mb height gradients, topographic influences (channeling in river valleys), the broad band of 850 mb isotachs (Appendix IVc), and two distinct regimes of low-level moisture. The latter will be discussed in Section 6.2. LJ4 flows across contours to lower heights, following the Rio Grande/Pecos river valleys (see Figure 3.1 for coarse topography). LJ1 flows northward east of the Edwards Plateau of central Texas under the influence of a strong surface pressure gradient (Appendix IVa). LJ1 played an important role in the formation and life cycle of the main AVE-V convective complex over Oklahoma. A shallow layer of moisture, with mixing ratios as high as 20 g kg^{-1} (Figure 6.3a,b), and trapped below a capping inversion at 900 mb, was transported northward from the Gulf of Mexico region by LJ1, and channeled into the storm formation region. Apart from its transportation role, this jet also figured in the dynamics of the situation, intersecting the upper jet (UJ1) over northeast Texas. At

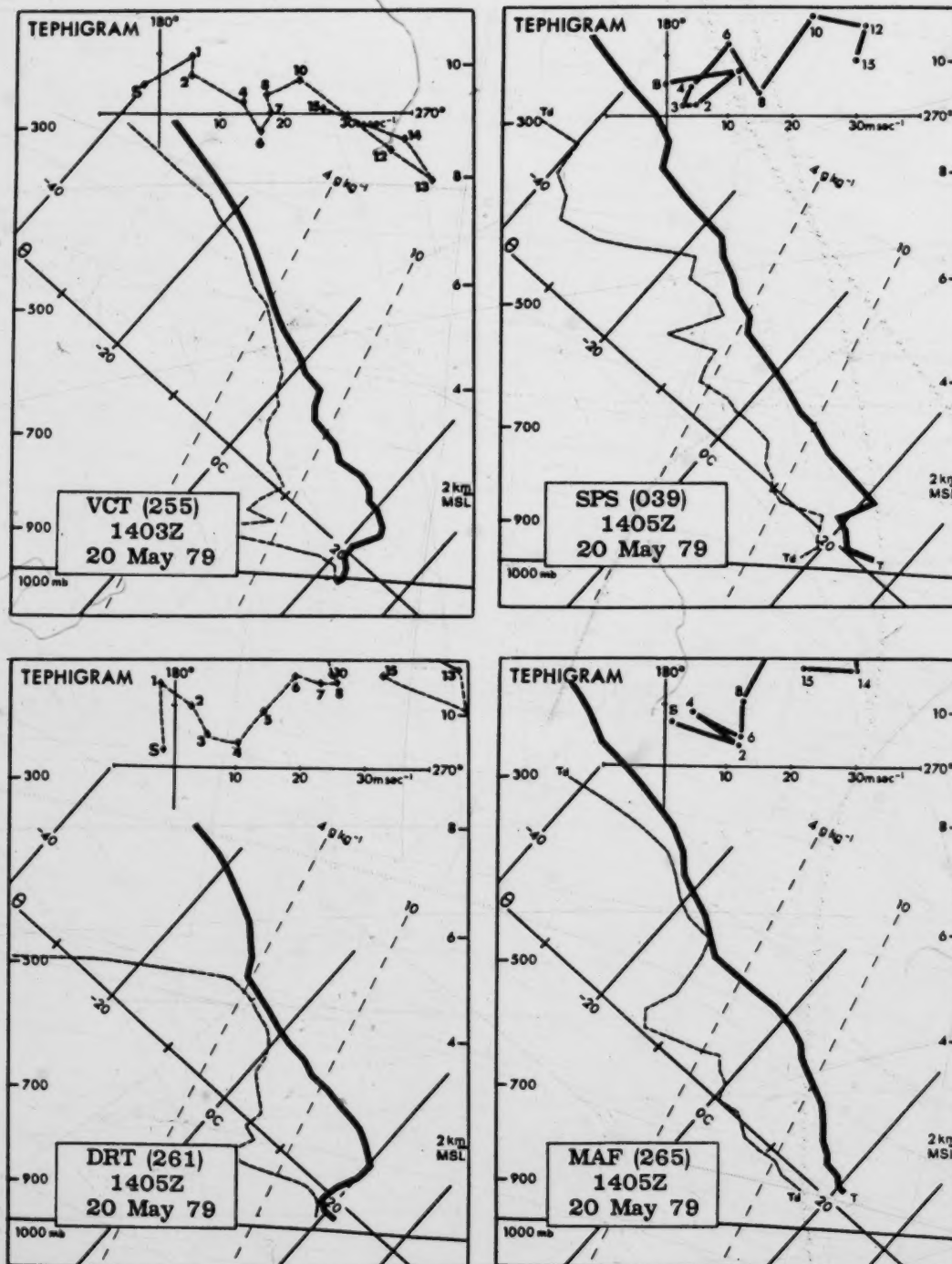


FIGURE 6.3: Soundings representative for low-level jets LJ1 (VCT and SPS) and LJ4 (DRT and MAF) during the pre-storm period, at 1400 UTC, 20 May, 1979.

202300, this intersection region coincided with the major storm over southeast Oklahoma (Figure 5.15), where there were radar echoes above 15 km. At the same time, LJ4 was transporting moisture towards the Midland (MAF) storm region. While the air approaching MAF had lower mixing ratios (Figure 6.3c), it was distributed through a slightly deeper mixed layer than that of LJ1 (Figure 6.3a). The next section discusses the moisture and sounding analyses in greater depth.

There were two other low-level jet streaks appearing at 202300. LJ2 seemed to have become an extension of LJ1 over southern Oklahoma, although a clear connection cannot be made because the region had light low-level winds (Appendix IVb). This region of relatively light winds resulted from the existing low-level convergence and ascent at this time (Figure 5.13), which was supporting the severe storms throughout southern Oklahoma.

LJ2 merged with LJ3 over the Texas Panhandle, which then appeared to converge with the moisture-laden LJ4 streak east of the Rockies along the west Texas/New Mexico border. This convergence of the two low-level jets appears to have been the main dynamic support for the regions of ascent and persistent storms over west Texas.

At the same time, the intersection of jets LJ4 and UJ1 over south-central Texas provides some dynamic support for the severe storms there. Radar echoes exceeding 16 km are noted over the latter region by 210235 (Figure 5.15). With the surface heating source disappearing after 202340 (sunset over the western regions of Texas/Oklahoma), it appears that these dynamics were the only remaining external source of energy.

By 211100, the jet UJ2 was over eastern Oklahoma, and was clearly interacting with the northern jet UJ3, near the Great Lakes. There appear to be two regions of favourable upper/lower jet coupling over southwest and northeast Texas, where the only significant active storms remained, with radar echoes to 13.0 and 11.8 km respectively (Figure 5.16). Other jet structures appear no longer favourable for convection, and with no remaining heat sources, all remaining storms were in advanced states of decay.

6.1.2 Kinetic Energy

While no detailed kinetic energy budget is being presented in this study, some aspects of the budget can be deduced by inspecting certain features of the upper winds. It was noted above that 300 mb wind speeds decreased significantly between 201100 and 202300. This effect was investigated further by analyzing the 12- and 24-hour changes in wind speeds at 300, 500, 700, and 850 mb (see Appendix IVc).

During the 12-hour period leading up to the mature AVE-V severe storm over Oklahoma, regions in the immediate vicinity of the storm underwent a loss of KE throughout the troposphere. The region south of the storm was also a KE sink at low levels (850 mb), but it became a source region at middle and upper levels. This pattern was the reverse northwest of the AVE-V storm, with a KE source at low levels, and generally a KE sink above. A significant sink existed at upper levels west to east across the storm. Some of the energy loss at middle and upper levels over and upstream of the storm, was regained 500-1000 km downstream to the northeast, where the greatest 24-hour energy gains

were at 700 and 300 mb and the winds increased by more than 12 m s^{-1} . This supports the argument that energy exchanges occurred between the two major synoptic systems to the southwest and northeast of the study area, in a manner similar to that given by Sechrist and Dutton (1970).

While it is not possible from this limited analysis to say that the convective storm actually extracted energy from the synoptic scale during its early development, the evidence certainly points in that direction. Some of these observations are verified by Fuelberg and Printy (1984), who computed time- and area-averaged kinetic energy budgets at both the synoptic and mesoscale for this AVE-V case. They showed a net loss of kinetic energy at all levels from the mesoscale network averaged through the period, and a small net gain in the 400-150 mb layer for the synoptic network (encompassing approximately the region shown in Figure 6.2). However, their energy budget was concentrated on the mature storm rather than the pre-storm stage.

The patterns and boundaries of sources and sinks over the severe storm region implies an energy exchange between the storm and synoptic scale energetics. For example, the 24-hour wind speed changes (201100 - 211100) show that the kinetic energy sink regions over Oklahoma, New Mexico, and west Texas coincided with the regions of convective storms. Also, the 700 mb pattern of KE loss during the first 12 hours is similar to the convective complex shape itself, although it is much larger in area. One cannot say that this loss was due to the storm blocking the flow around it, nor is it certain what significance this has to any energy interactions between the synoptic flow and the storm complex at the mesoscale.

The 850 mb energy gain on the northwest side of the AVE-V complex is most likely due to synoptic processes, since the northerly jet streak over that region at 202300 (LJ3 in Figure 6.2) had started to show up over Nebraska as early as 201100. On the other hand, the KE sink at 850 mb south of the storm, if not due to advection, would be related to the gain aloft over the same region; that is, this energy may be necessary to transport air aloft where it could release sensible and latent heat. The energy decreases may thus be considered as a contribution of environmental kinetic energy to the convective storm.

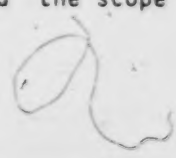
In this limited energy analysis, one can only suspect that storms consumed some energy transferred down from the synoptic scale, while feeding back some of the energy to the larger scale downstream through latent heat release. If indeed the larger scale flow is primarily responsible for the mesoscale vertical circulations through a downscale transfer of energy, then any resulting low-level convergence would lead to cross-isobaric flow. In turn, this would cause mesoscale regions of baroclinicity, and a subsequent conversion of potential energy to kinetic energy via the processes described in Section 2.6. Since the mesoscale ascent also leads to destabilization of the airmass, especially during initial stages of convection, this energy conversion would take place in the inflow region of a resulting storm complex.

Further theoretical support for this is deduced from the dynamics of synoptic scale flow. Recall that the principal dissipation of large scale kinetic energy is not directly through molecular friction, but rather through smaller intermediate eddies (Lorenz, 1955). At ever smaller scales, the kinetic energy of these eddies eventually is dissipated by molecular friction. This appears to be the reason why

"for synoptic scale numerical models where turbulence is approximately two-dimensional, microscale (sub-grid scale) processes are not capable of influencing the large scale flow significantly" (Tennekes, 1978), and one can parameterize friction effects out to several days.

Based on the latter observation, one may deduce that the feedback of energy from convective storms to the synoptic scale flow does not normally surpass the storm's average consumption of large scale energy in total, since these storms also dissipate vast amounts of energy. Also, this feedback does not greatly influence the larger scale system. Furthermore, the evidence presented thus far makes it fairly clear that synoptic scale processes can greatly influence the mesoscale storm environment, especially during the pre-storm period.

In spite of this, it is rather perplexing to the author that the majority of studies alluding to severe storm scale interactions, concentrate on the mature storm stages and possible feedback mechanisms, and tend to ignore what are probably at least equally important downscale processes during the pre-storm period. A more detailed space- and time-resolved energy budget analysis at different scales of motion, using data from one or more mesoscale networks nested within the larger synoptic network, would be a logical step for future research into energetics. One could then examine source and sink terms at the two scales, and at different times and locations. The comparison might provide the answers necessary to confirm the scale interactions suggested above. However, this is beyond the scope of the present study.



6.2 SOUNDING ANALYSES

A detailed analysis of tephigram soundings was carried out in order to obtain estimates of the thermodynamic potential (static stability) of storm regions. The purpose of this analysis was to examine possible forecasting parameters from the pre-storm period; that is, prior to the first radar echoes at 201830. Results are therefore presented for only the 201400 and 201700 soundings. All soundings were analysed subjectively, employing the manual techniques practised by forecasters for many years (Means, 1952; Showalter, 1953; Miller, 1972; Renick and Maxwell, 1978; Strong, 1979).

A brief review of the procedures used and the derived sounding parameters follows. First, a mean mixing ratio was estimated within the boundary layer of the sounding. The boundary layer was determined as the layer below a capping inversion base, the latter generally being within the 850-800 mb layer for the AVE-V case. For soundings without such a capping inversion, an average mixing ratio of the lowest 75 mb of the sounding was used (Strong, 1979). An estimate of cloud base, or level of free convection (LFC), was then chosen to be the pressure at the intersection of this average mixing ratio and the temperature sounding. The wet-bulb potential temperature, Θ_w at the LFC was the next parameter calculated, representing the maximum, in-cloud pseudo-adiabat. An indication of the degree of static instability was then estimated, using the lifted index at 500 mb (LI500). This is the difference between the cloud parcel temperature, obtained from Θ_w , and the sounding temperature at 500 mb. The main difference between LI500 and the Showalter Index (Showalter, 1953) or the Best Lifted Index of Fujita et al. (1970), is in the sign, where Canadian forecasters,

contrary to their U.S. counterparts, prefer to use positive numbers for instability. Area average values and cross-correlation tables are provided in Appendix V for the various sounding parameters, for both the mesoscale and synoptic domains at 201400 and 201700.

6.2.1 Sounding Moisture

The sounding-derived analyses of mixing ratio, representative for convection at 201400 and 201700, are presented in Figure 6.4. The diamond symbols indicate the locations of first radar echoes during the first hour of storm development after 201830. Clearly, the main storms did not form where the available moisture was at a maximum. The hourly mesoscale surface and 850 mb analyses of Appendix IVa and Figure 6.2 respectively, suggest that these storms formed along a sharp dryline, where higher humidity was being pumped into the southeast sector of

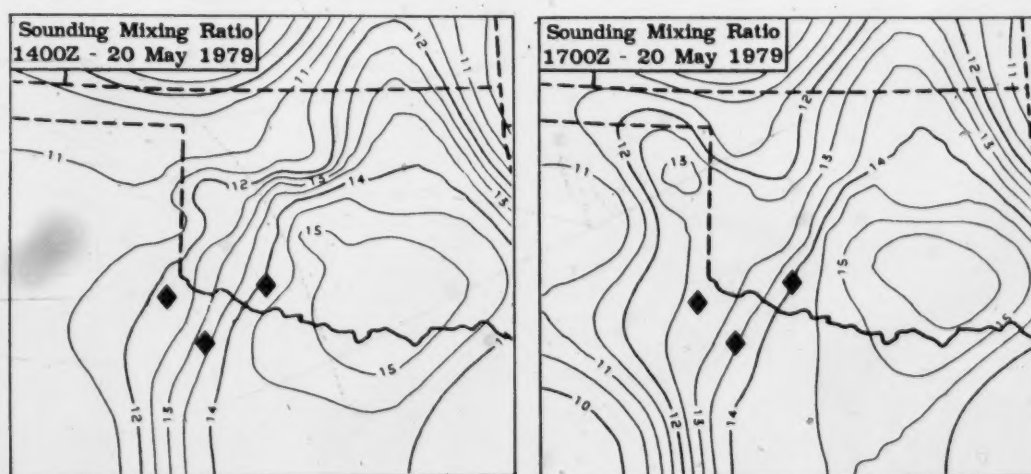


FIGURE 6.4: Sounding-derived analyses of mixing ratio (g/kg) representative for convection, at 1400 and 1700 UTC, 20 May, 1979. Diamond symbols indicate locations of first radar echoes, generally after 1830 UTC.

potential storm regions. This dryline feature appears to be common to storms in the lee of the Rockies throughout North America, including Alberta (Strong, 1981).

The distribution of mixing ratio did not change significantly over the three hours between 201400 and 201700, except for the organization of lines of moisture north and northwest from the mixing ratio maximum over south-central Oklahoma. These two 'tongues' of moisture correspond to the two main centres of action of the AVE-V complex. A similar dual pattern can be noted in the 700 mb vertical motion field at 201830 (Figure 5.9). The two moisture tongues are separated by the trough of drier air pushing southward over western Oklahoma, the latter being related to the main centre of subsidence over the region during the pre-storm period (Figure 5.9). Based on this moisture pattern, the AVE-V storms appear to be unrelated to the Midland storms (on the southwest edge of this map) where the mixing ratios were decidedly lower ($< 10 \text{ g kg}^{-1}$). These storms apparently formed under different circumstances.

6.2.2 Surface Temperatures for Convection

Convective temperature is defined as the surface temperature required to achieve the level of free convection (LFC) and therefore convective clouds, based only on the static stability of a given sounding. The analysis of predicted convective temperature (T_c), shown in Figure 6.5, is revealing in two ways. First, it illustrates how misleading it can be to use a single sounding to determine this quantity over the synoptic domain with only synoptic scale soundings. For

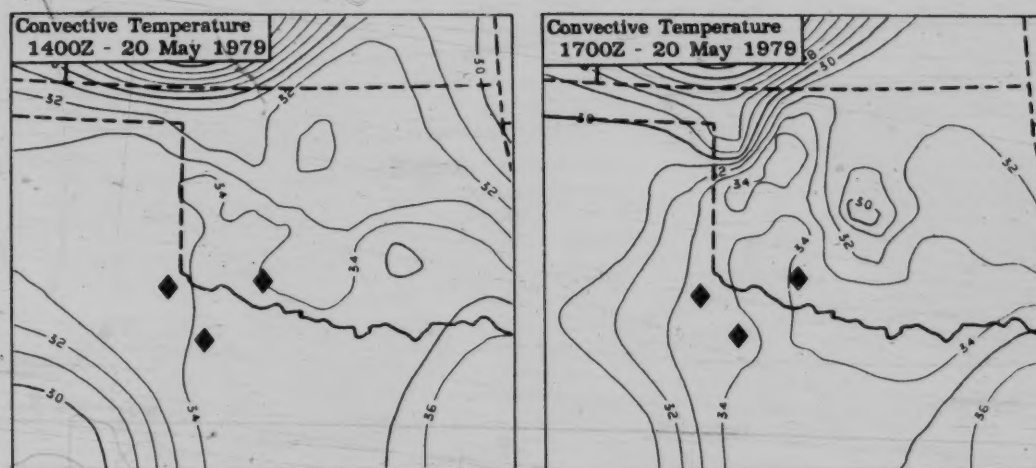


FIGURE 6.5: Sounding predictions of convective temperature ($^{\circ}\text{C}$) at 1400 and 1700 UTC, 20 May, 1979. Diamond symbols indicate locations of first radar echoes.

example, the mesoscale gradients of T_c shown here are particularly important between Oklahoma City (OKC) and Midland (MAF) (indicated by open triangles), the region where the major storms formed. Across this region, T_c increases to $>34^{\circ}\text{C}$, then decreases again. Without the additional mesoscale sites over southwest Oklahoma, the higher potential (high T_c) between the two synoptic sites would be unknown. Also, for the region between OKC and Dodge City (DDC) to the northwest, T_c decreases rapidly only near DDC.

Second, the pattern and values of predicted convective temperature, and of sounding mixing ratio, do not change significantly during the pre-storm period from 201400 to 201700. These are the two parameters required for forecasting convective clouds from parcel theory, and this suggests that if a forecaster is limited to one or two sounding sites, little, if anything, is gained by having soundings at three-hour

intervals. In this case, what he really needs is more spatial resolution, so that mesoscale features can be observed. The two lines of maximum T_c in Figure 6.5 also suggest two main potential regions for convection, over eastern and western Oklahoma. However, no similarity in patterns between the Oklahoma (AVE-V) and west Texas (MAF) storms is evident in these analysis.

Figure 6.6 shows the spatial and temporal mesoscale changes in surface temperatures (T_s), as determined from the 201400 and 201700 soundings only. Figure 6.7 shows the surface temperature increases ($T_c - T_s$) required to realize convective temperature at the same times. Thus, for the main (AVE-V) storm at least, convection occurred first over the region where the surface temperatures were highest several hours before, and, not surprisingly, where the least amount of heating was required to achieve convective temperature. In the latter respect, the west Texas (MAF) storms were no different. The occurrence of

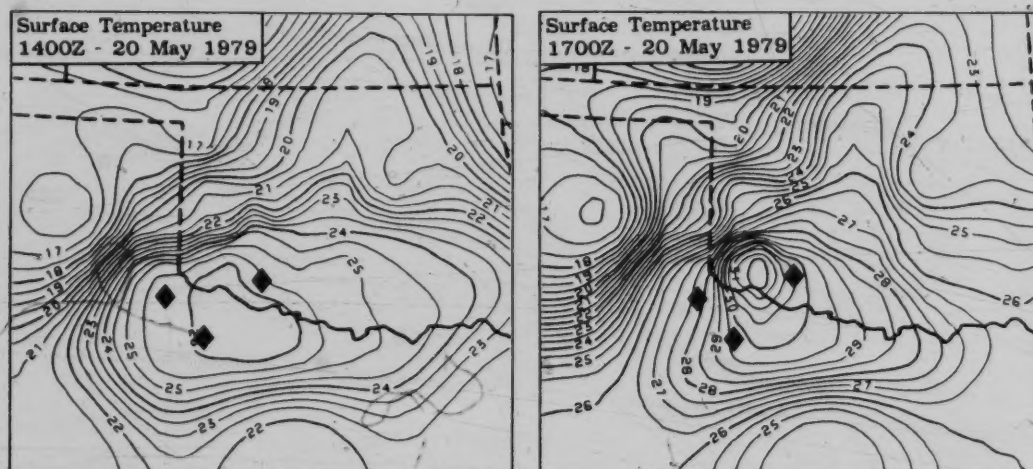


FIGURE 6.6: Surface temperature fields ($^{\circ}\text{C}$) from soundings at 1400 and 1700 UTC, 20 May, 1979. Diamond symbols indicate locations of first radar echoes.

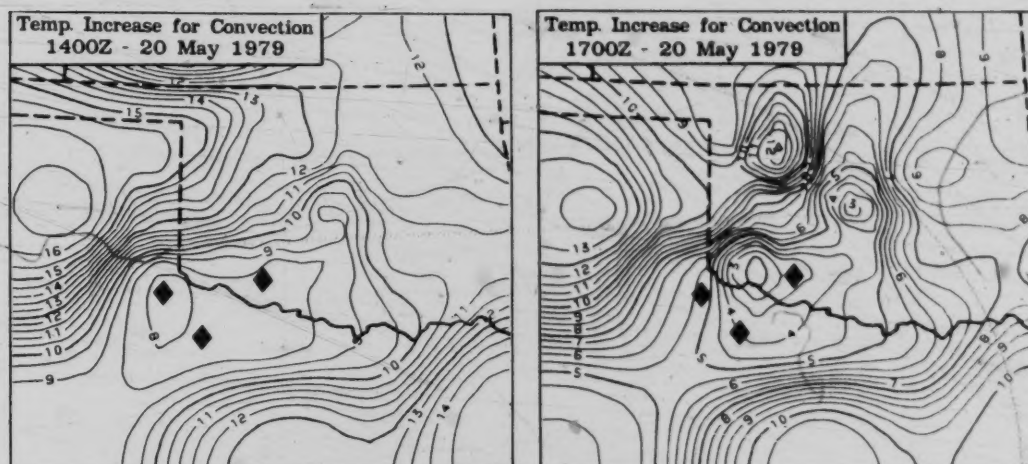


FIGURE 6.7: Surface temperature ($^{\circ}\text{C}$) increases necessary to realize convective temperature at 1400 and 1700 UTC, 20 May, 1979. Diamond symbols indicate locations of first radar echoes.

cloudiness would obviously have an effect here, but one also should not overlook the influence of a spatially-varying surface albedo.

We next examine actual surface temperature increases between 201400 and 201700, as depicted in Figure 6.8. Both the Midland and AVE-V storms were initiated near two regions which experienced the most rapid pre-storm temperature increases. However, it appears that initial AVE-V storm cells formed about 75 km southwest of the centre of maximum heating over southwest Oklahoma. At 201700, these two regions (Figure 6.8) also required the least temperature increase to attain convective temperature (Figure 6.7). Again, the spatial variations in Figure 6.8 probably result from variations in cloud cover and surface albedo, although one cannot overlook the effects of surface moisture. If much of the insolation is used to evaporate surface water, then the temperature will not rise so rapidly. The effect is counteracted by

evaporation increasing the surface moisture supply (mixing ratio), thereby lowering the temperature required for convection (T_c).

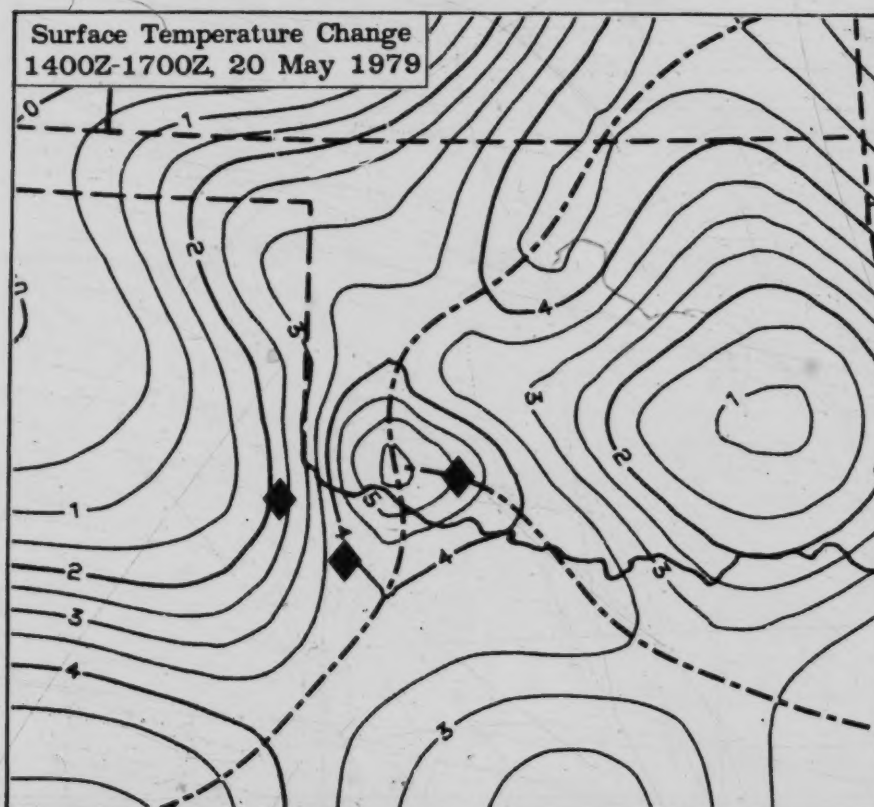


FIGURE 6.8: Surface temperature changes ($^{\circ}\text{C}$) between soundings, 14-1700 UTC, 20 May, 1979. Diamond symbols indicate locations of first radar echoes, broken lines indicate maximum heating.

Meanwhile, the hourly surface analyses (Appendix IVa) show that maximum surface temperatures were attained throughout Oklahoma and the immediate adjacent regions between 201700 and 201900, and that the overall maxima varied little from the surface temperatures at 201700 (Figure 6.6). Therefore, it can be deduced that maximum temperatures over the storm formation regions fell short of the sounding-predicted

convective temperatures by at least 2-4 °C; that is, by approximately the amounts indicated in Figure 6.7 at 201700.

The effects of surface heating, along with topographic influences, certainly cannot be overlooked as primary sub-mesoscale influences on convective storms. Nevertheless, the occurrence of severe convection in this case, despite the failure to achieve T_c , was primarily a result of ascent and adiabatic cooling of the capping inversion prior to convection. An important factor to remember for forecasting considerations, is that sub-grid, or sub-mesoscale effects, such as topographic and surface albedo variations, are 'fixed' for all intents and purposes, and are deterministic in any modeling effort.

6.2.3 Convective Instability

The level of free convection (LFC) is estimated graphically on a tephigram from the predicted T_c (Figure 6.5) and the sounding mixing ratio (Figure 6.4), and represents convective cloud bases. It would appear from Figure 6.9 at 201700, that the convective cloud bases were initially (201830) near 750 mb over southwest Oklahoma, but that these dropped to 800 mb (or lower) for new cells which formed further east over south-central Oklahoma 1/2-1 hour later. Cloud base temperatures (Figure 6.10) were generally 13-16 °C, but may have been as high as 18°C over southeast Oklahoma. Thus, storms do not necessarily form where the potential cloud base temperature is highest.

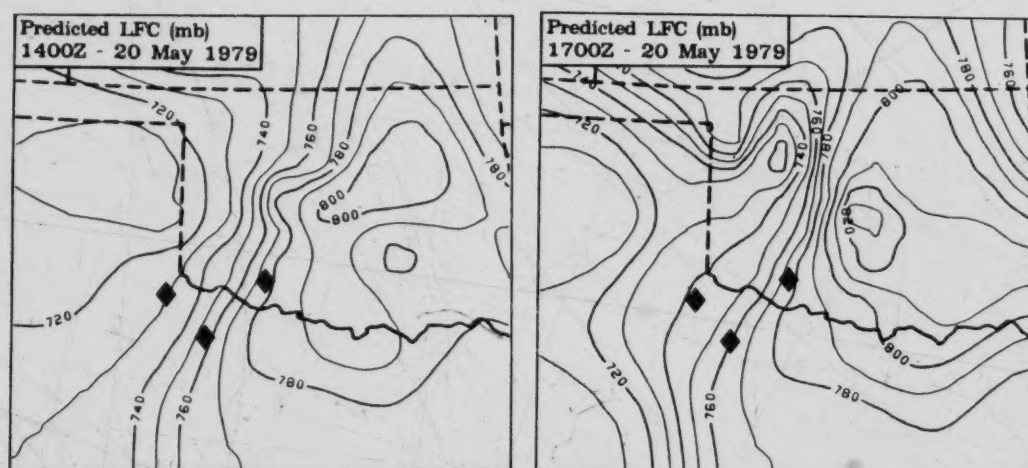


FIGURE 6.9: Predicted cloud bases (LFC, mb), using the mixing ratios of Figure 6.4 and the convective temperatures of Figure 6.5, for 1400 and 1700 UTC, 20 May, 1979. Diamond symbols indicate locations of first radar echoes.

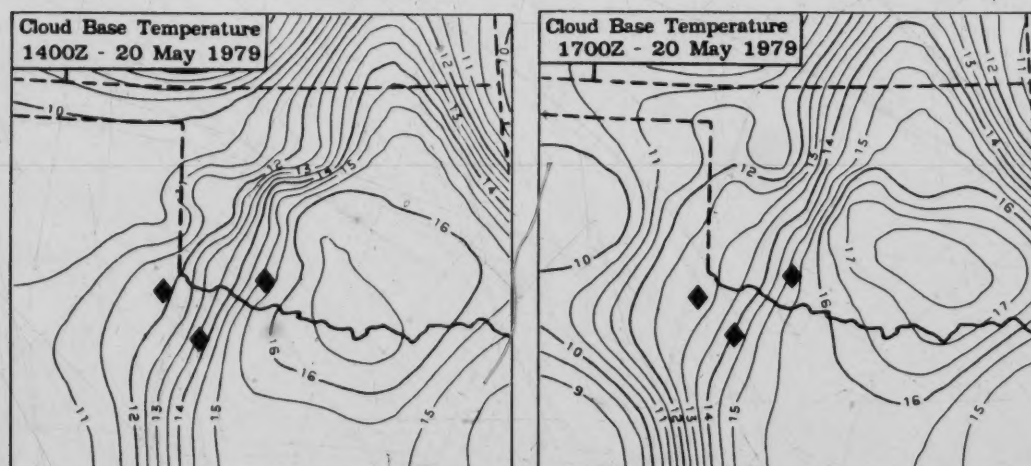


FIGURE 6.10: Predicted cloud base temperatures ($^{\circ}\text{C}$), using the mixing ratios of Figure 6.4 and convective temperatures of Figure 6.5, for 1400 and 1700 UTC, 20 May, 1979. Diamond symbols indicate locations of first radar echoes.

Over west Texas, the LFC was closer to 700 mb in the drier air, while cloud base temperatures were $<10^{\circ}\text{C}$. These storms were more

comparable to central Alberta severe storm cases, where typical convective cloud bases are 750 mb and cloud base temperatures are 5-10°C. Both regions have a range of the Rocky Mountains just to the west, with their respective convective storms forming in the lee of these mountains.

Analyses of the sounding-derived cloud adiabats (Theta-W) and lifted index (LI500) appear in Figures 6.11 and 6.12 respectively. Storms formed upstream, west and northwest of the centres of maximum Theta-W and Lifted Index. A typical Theta-W value was 24°C for the AVE-V complex, and <22°C for the Midland storms. Lifted Index correlates highly with maximum updraft velocity in a convective cell, and therefore to the size of hail it is able to sustain (Chisholm, 1973). The variation of the index across relatively short distances gives some idea of the uncertainty involved in predicting the intensity of convective weather based on single soundings of the synoptic network.

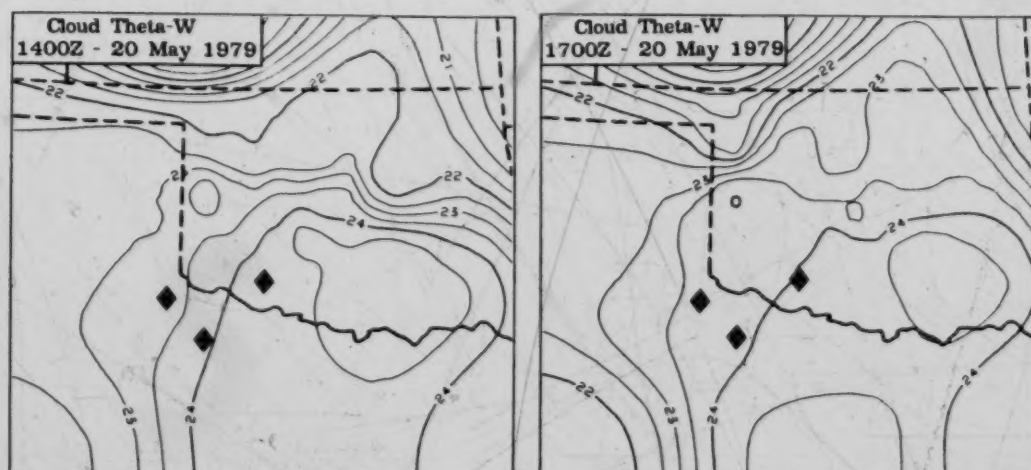


FIGURE 6.11: Predicted cloud adiabats (Theta-W, in °C), using the mixing ratios of Figure 6.4 and the convective temperatures of Figure 6.5, for 1400 and 1700 UTC, 20 May, 1979. Diamond symbols indicate locations of first radar echoes.

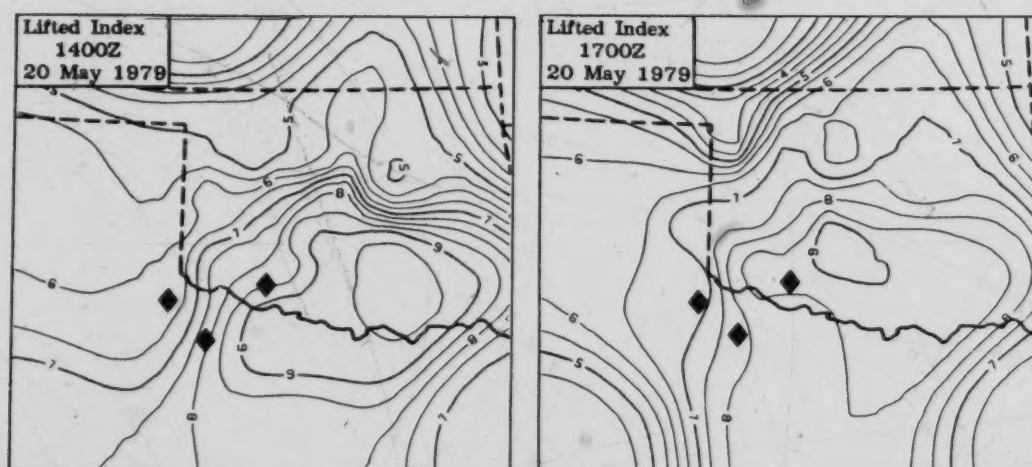


FIGURE 6.12: Predicted parcel lifted index (LI500, in °C), using the mixing ratios of Figure 6.4 and convective temperatures of Figure 6.5, for 1400 and 1700 UTC, 20 May, 1979. Diamond symbols indicate locations of first radar echoes.

6.3 MESOSCALE MOISTURE SOURCES

The KE source region was inferred (Sections 5.3 and 6.1.2) to be southwest of the AVE-V complex. The source of the energy derived from diabatic surface heating is of course local to each storm. It appears that much of the kinetic energy input to storms goes toward the vertical transport of moisture, apart from that lost to frictional or eddy dissipation. But this moisture itself is a source of latent heat energy. Therefore, the consideration of energy sources cannot be complete without investigating source regions for moisture.

6.3.1 Surface Analysis of Mixing Ratios

Figure 6.13 shows the changes in surface mixing ratio over the synoptic space and time scale, from 201200 to 210000. The 850 mb jet streaks of Figure 6.2 are superimposed. It seems clear that the main jet streak was drawing moisture northward from the Gulf of Mexico. Initially, this increases the depth of moisture within the boundary layer, and later it is required to feed the growing convective storms. Little net increase of surface moisture is therefore noted between the mixing ratio analyses of 201200 and 202400.

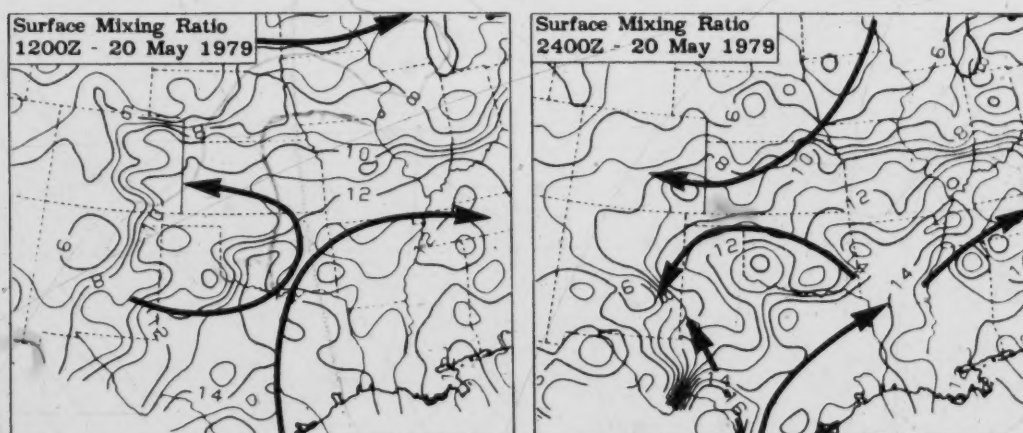


FIGURE 6.13: Surface mixing ratios (g/kg) at 1200 and 2400 UTC, 20 May, 1979, with 850 mb maximum wind streaks of Figure 6.2 super-imposed.

The soundings of Figure 6.14 for Victoria (VCT), Texas, situated near the source region, indicate that between 201100 and 201400, this moisture from the Gulf of Mexico was quite shallow, but that it increased in depth with time. The soundings of Figure 6.15 for Longview

(GGG), Texas, show a somewhat deeper layer of moisture initially, which also increased with time. Soundings over the mesoscale network near where the jets converge, reveal yet deeper layers of moisture, with mixing ratios exceeding 15 g kg^{-1} (e.g., see Figures 6.23, 6.24). It should be understood that this moisture remained below the inversion in the lowest 50-100 mb of the sounding, until convection took place. The wide extent of this shallow layer is evident in the dew point depression analyses of Figure 5.22, which show a very dry airmass at 850 mb just above the inversion. The role of the jet streak in such instances is to channel the moisture, in under the capping lid, and, through convergence and low-level ascent, organize it into relatively deep tongues of moisture.

The western-most jet streak (LJ2) at 201200 (Figure 6.2) also appears to have played a role in this moisture channeling over Oklahoma. A second southerly low-level jet (LJ4), later advected moisture northwest along the Rio Grande and Pecos River valleys (Figure 6.13b). While the surface divergence field was not computed for this time, the orientation of the jet streaks converging over the Midland region, suggests that this region experienced the strongest low-level convergence at 210000 (compared with the 201400 pattern of computed divergence in Figure 5.6f). Thus, the Midland storms would likely have been more severe, had the jet streak LJ4 been transporting moisture into the region for a longer period. Meanwhile, the (AVE-V) storm complex over Oklahoma, in addition to losing its heat source after 210000, also lost its source of moisture as the jet streaks became reconfigured.

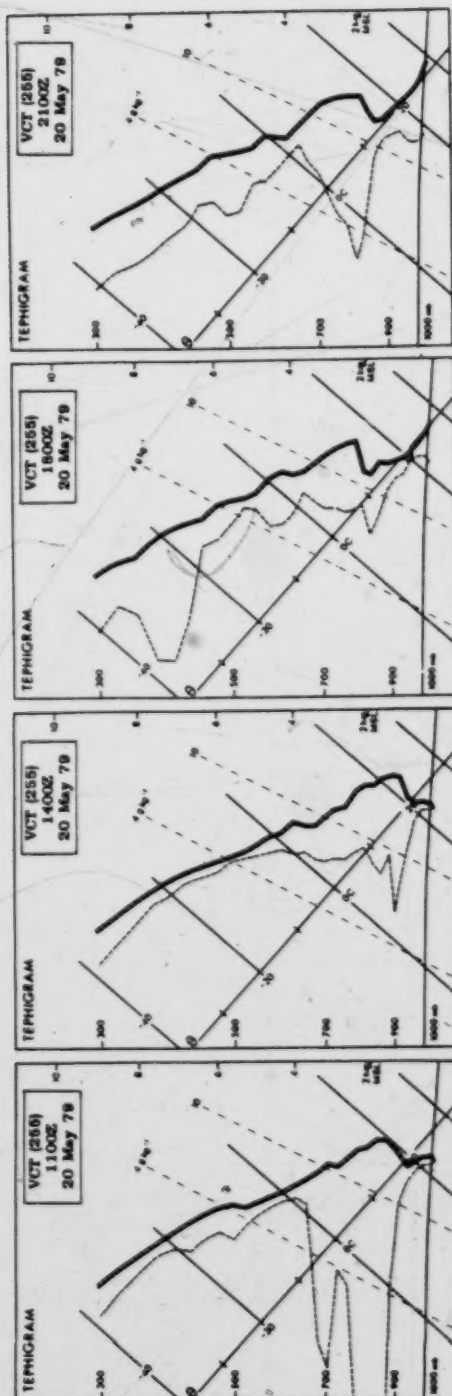


FIGURE 6.14: Sequential tephigram soundings for Victoria, southeast Texas (VCT).

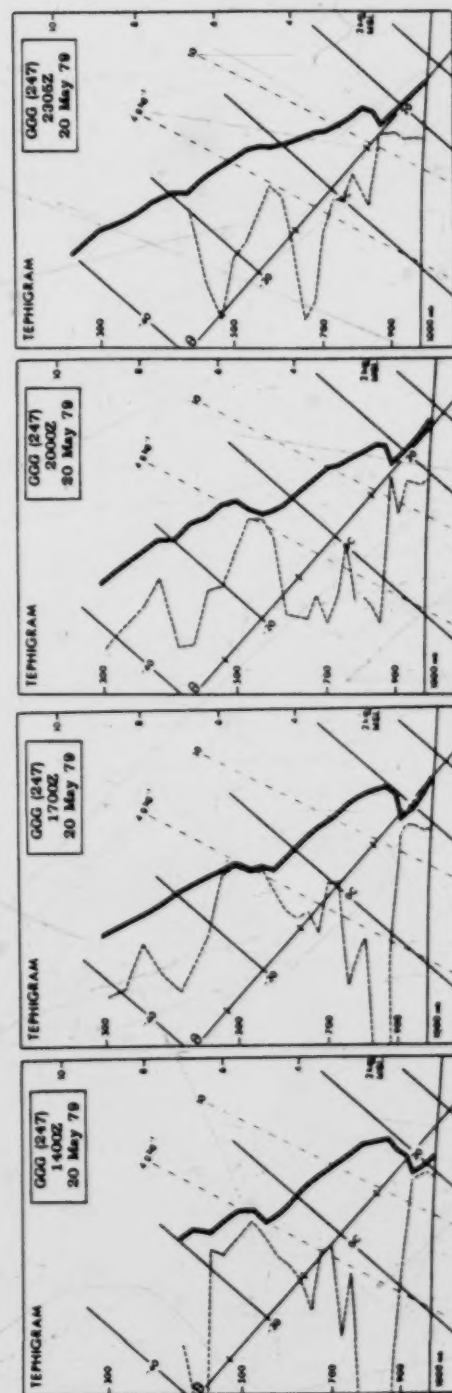


FIGURE 6.15: Sequential tephigram soundings for Longview, east Texas (GGG).

6.3.2 Comparison of Surface and Sounding-derived Mixing Ratio Fields

The usual approach to determine available moisture for storms from sounding information, is to estimate an average mixing ratio within the boundary layer, in order to account for convective mixing. The values of mixing ratios as shown in Figure 6.4 for 201400 and 201700, were averages of the lowest 50-75 mb. To examine how important this mixing is over the region, Figure 6.16 is a comparative analysis for the same times, using unmixed values of mixing ratio from the surface data alone. That is, Figure 6.4 is based on vertically averaged mixing ratios from the sounding sites of Figure 4.2, while Figure 6.16 uses unmixed surface data from the sites of Figure 4.3. Therefore, Figures 6.4 and 6.16 have comparable resolution only over the mesoscale network of western Oklahoma. Beyond this, the sounding sites are spaced 370 km on the average, while surface sites are typically 130 km apart.

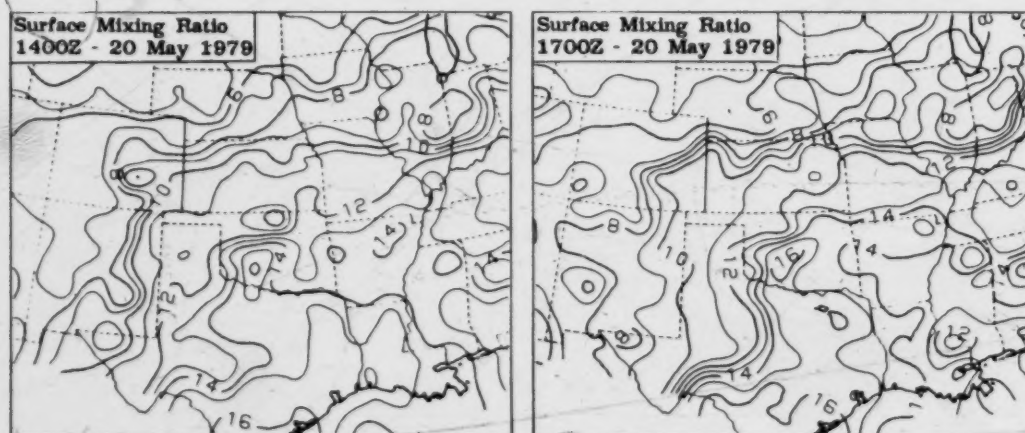


FIGURE 6.16: Surface mixing ratio field (g/kg) at 1400 and 1700 UTC, 20 May, 1979.

Some of the differences in the two pairs of patterns are due simply to this difference in spatial resolution. For example, the surface data (Figure 6.16) indicate a relative dry region over eastern Oklahoma at 201400, with mixing ratios $< 13 \text{ g kg}^{-1}$, where the interpolated sounding values (Figure 6.4) range from $14\text{--}16 \text{ g kg}^{-1}$. Other differences stem from the fact that sounding values, obtained by mixing the lowest layer, can either increase or decrease the mean surface value. In other words, moisture may be either minimized or maximized at the surface. The latter differences are believed to be a direct result of the capping inversion, to be discussed further in Section 6.4.

The surface analysis at each time indicates the moisture maximum to be closer to, although still east of where storms eventually formed. This is probably more correct than the sounding analysis, since there were no sounding sites over the eastern third of Oklahoma (Figure 4.2), the closest being Little Rock, Arkansas. The sounding values at 201700 more correctly indicate a dry zone appearing north of the storm formation region (northwest Oklahoma), which remained storm-free, at least during the early stages of these storms. Both sets of data indicate a strong gradient of mixing ratio over and south of the storm formation region, sometimes referred to as the leading edge of a dry line.

It is not clear from this comparison, that sounding-derived mixing ratios over the mesoscale network were any better in this regard than the less dense surface site data. Outside the mesoscale network, however, one would be wise to take advantage of the greater density of surface data. Using either data set, the inferences of the last section remain valid here. That is, the main storms formed upwind from the

centre of maximum moisture, where low-level winds were from the southwest (Figure 6.2). Above all, this comparison illustrates the value of surface information, too often under-used in diagnostic studies, and it emphasizes the importance of increasing the spatial resolution of surface data in mesoscale experiments wherever possible.

6.4 THE CAPPING LID DURING AVE-V

The notion of a capping lid, as originally defined by Beebe and Bates (1955), was introduced in Section 1.3.3. Chapter 2 reviewed some theories of atmospheric energetics, and how mesoscale motions might be related to the larger scale through eddy dissipation of kinetic energy. An extended hypothesis of the pre-storm formation and breakdown of the capping lid, and of its role in Alberta hailstorms, was then presented in Section 3.2. This explanation encompassed synoptic scale motions and the concept of mesoscale circulations, complete with their own mesoscale fields of vertical motion and low-level convergence of moisture.

In Chapters 4 and 5, these regions of mesoscale circulations, particularly the vertical motion fields, were shown to exist prior to organized convection during the AVE-V data period, and, within the limits of available data, they were shown to be related to both synoptic dynamics and the observed convective complexes. The preceding sections of this chapter have examined the role of jet streaks during AVE-V, and their relation with the boundary layer moisture field. It now remains to relate together all of these factors in terms of the convective storms of the AVE-V data period.

6.4.1 The Role of the Capping Lid During AVE-V

As in many, perhaps most, severe convective storm cases, the capping inversion (or lid) is a common pre-storm feature of the AVE-V sounding data set. Its identification depends not only on inversions of dry bulb temperature, but also inversions of "potential temperature". For this reason, capping "lid" is often preferred as a more descriptive term. Also, discussions concerning such lids, usually refer only to those inversions or stable layers capped between 900 and 700 mb, depending on elevation above sea level¹⁵. In the literature, we find that 800 mb is the average level for the pre-storm lid (e.g., Fawbush and Miller, 1954). This is also true for higher plains regions, including Alberta (Strong, 1982). A lid capped at 800 mb provides a capping inversion depth for trapping moisture of 150-200 mb for the Oklahoma-Texas region, but only 100 mb for Alberta. The intensity of the inversion and the depth of trapped moisture appear to be major factors in determining the horizontal extent of the lid influence. In this study, another lid factor is the size of pre-existing atmospheric circulations, which to a large extent, also determines the eventual size of convective complexes.

The ultimate role of the lid in severe storm situations is simply to suppress the release of convective instability, while latent instability increases due to the build-up of moisture below the inversion, as Figure 6.15 shows. This combination of moisture, jet streaks which can channel the moisture into or moist tongues, and the

¹⁵ The term capping inversion, or lid, as used here, does not include the more common nocturnal inversion, which results primarily from surface radiation, and is usually capped at less than 50 mb (< 500 meters) above the ground.

capping lid, completes the three-dimensional picture of moisture convergence. As pointed out by Carlson et al. (1983), intense convection can then occur where the moist air flows out from beneath the lid, the process called "underrunning". Convection will occur preferentially on the upwind, cold, dry edge of the lid region, where there is usually pre-existing ascent to weaken the lid and allow the release of convection. Mesoscale ascent approaches the lid region from upstream (with respect to mid-level winds), which is why the cold side of the lid edge is generally affected first. The extent to which these two effects combined for the the main convective storms during AVE-V will now be investigated briefly.

6.4.2 Causes for the Capping Lid during AVE-V

The causes of a capping lid were postulated in 3.2. Carlson et al. (1983) have suggested that the lid frequently originates from "the differential advection of a hot, dry mixed layer from the elevated plateau of Mexico (or the southwest United States) over a cooler, moister layer advected northward ahead of an upper trough". However, this explanation alone would not explain the occurrence of similar lids over other regions of the High Plains such as Alberta, where there are no such high elevated plateaus, only a west-east series of mountains and valleys. While their explanation cannot be discounted in the AVE-V case presented here, adiabatic warming and drying at mid-levels due to air subsiding from the Mexican plateau to the south, and from the mountains to the west, in addition to large-scale subsidence preceding an upper ridge, appear to be more important causes for the lid in this instance.

6.4.3 Regional Extent of the Capping Lid During AVE-V

Pre-storm soundings discussed in Sections 5.4.2, 6.1.1, and 6.3.1, exhibited marked capping inversions over the mesoscale network (ADA and CSM), and as far west as Amarillo (AMA), over the Texas Panhandle (Figure 6.17). These suit the conceptual model of elevated mixing overrunning a moist layer, as presented by Carlson et al. (1983). Closer to the Mexican plateau, however, Figure 6.18 shows that the capping lid was absent from pre-storm soundings at Midland (MAF), other than a nocturnal inversion at 201100, yet was very prominent at Del Rio (DRT), Texas, and downstream at Stephenville (SEP). It was also conspicuous further east near the coast at Victoria (Figure 6.14), and downstream at Longview (GGG) (Figure 6.15). Looking further afield from the Mexican plateau, Figure 6.19 shows that the lid was observed as far east as Lake Charles, Louisiana (LCH), with a hint of it even further east at Jackson, Mississippi (JAN), and as far north as Monett, Missouri (UMN). This information was combined to produce the schematic of Figure

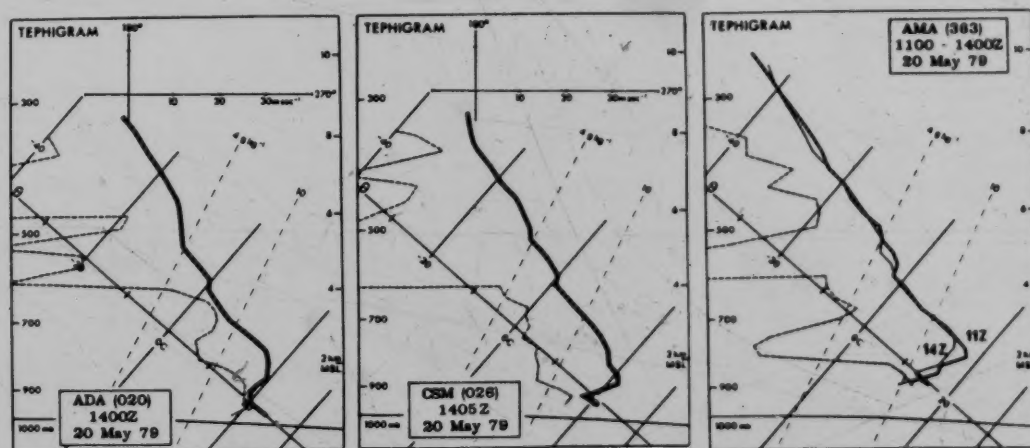


FIGURE 6.17: Pre-storm capping lid soundings for Ada (ADA) and Clinton-Sherman (CSM), Oklahoma, and Amarillo (AMA), Texas, 20 May, 1979.

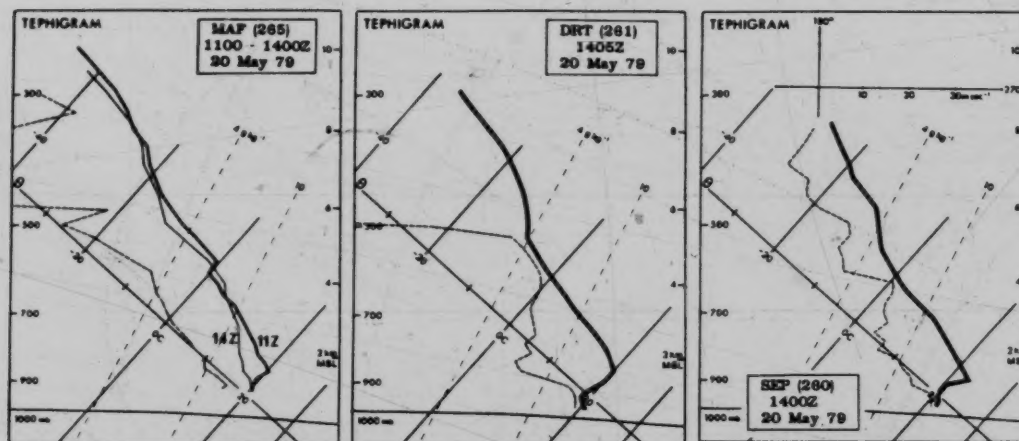


FIGURE 6.18: Pre-storm soundings for Midland (MAF), Del Rio (DRT), and Stephenville (SEP), Texas, 20 May, 1979.

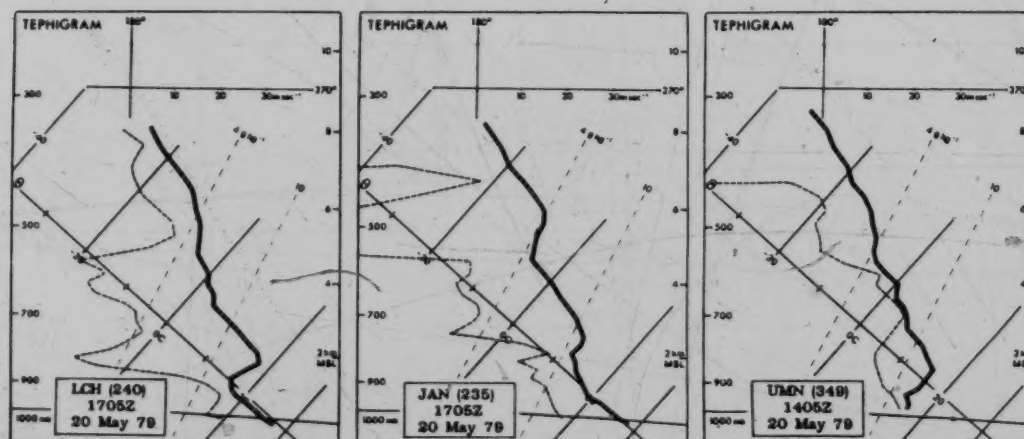


FIGURE 6.19: Pre-storm capping lid soundings for Lake Charles (LCH), Louisiana, Jackson (JAN), Mississippi, and Monett (UMN), Missouri, 20 May, 1979.

6.20, which shows the approximate regional extent of the pre-storm capping lid. In view of the southerly to southwesterly winds aloft (Figure 6.2), and the location of the Mexican Plateau, processes in addition to elevated mixing appear to have been at work.

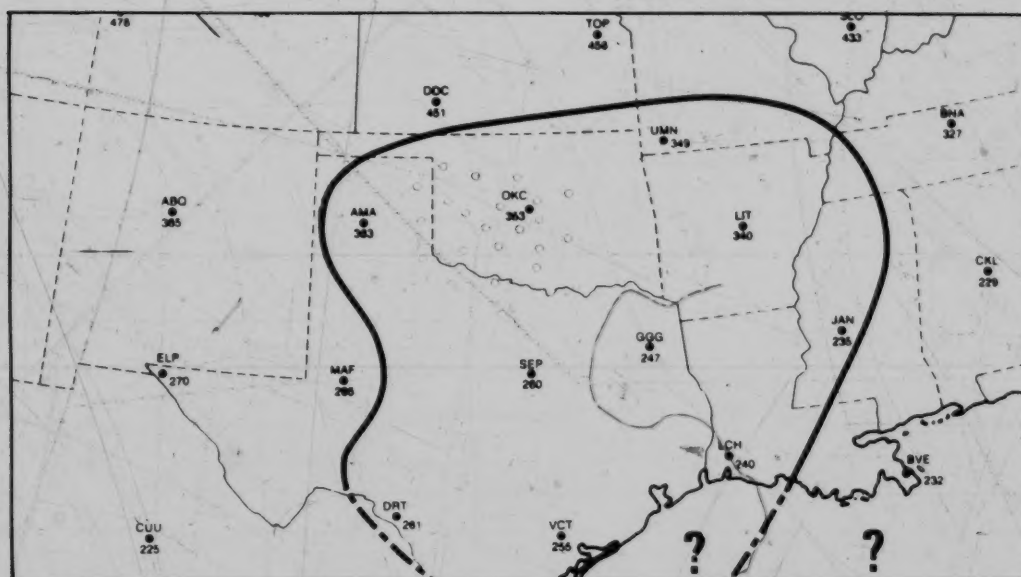


FIGURE 6.20: Approximate regional extent of the AVE-V pre-storm capping lid, 20 May, 1979. Its boundaries over the Gulf of Mexico are unknown.

To help reduce the subjectivity of defining capping lids in Figure 6.20, and to further identify optimum conditions for the severe storms of this period, the regional variations in height, temperature, and depth of the lid were examined in some detail. Figure 6.21 shows pre-storm analyses of the pressure and temperature at the top of the capping lid, and its depth from the lid top to ground level. Between 201400 and 201700, the general rise and cooling of the lid cap is observed. Severe storms formed over regions where the lid was capped near 800 mb an hour or so earlier, and where the lid temperature exceeded 17°C . During the three-hour pre-storm interval, the lid was lifted by an average of 30 mb and cooled an average of 3°C due to adiabatic ascent over the region. Storm initiation occurred where the lid was of medium depth, between 120 and 160 mb (Figure 6.21c).

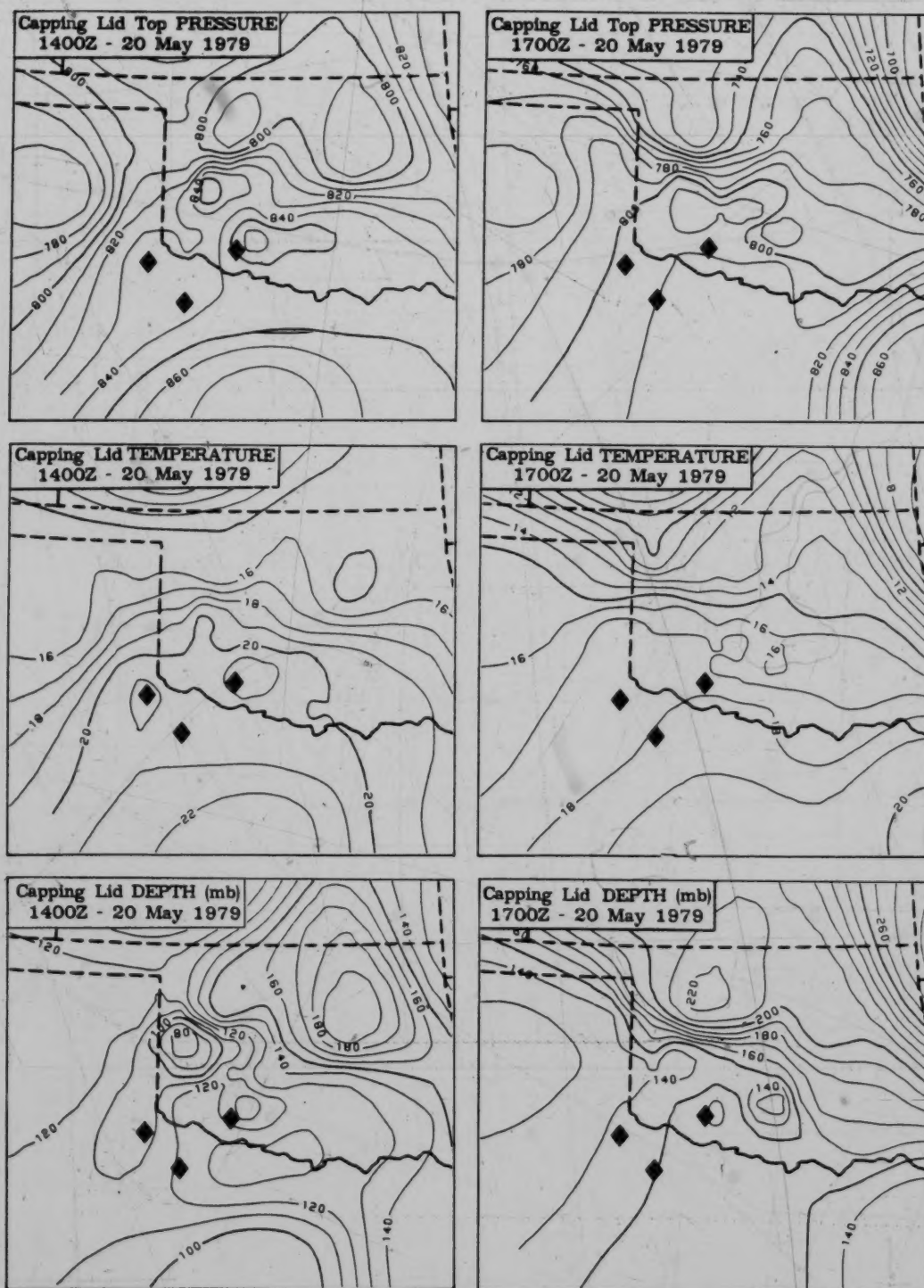


FIGURE 6.21: (a) Pressure (mb) and (b) temperature ($^{\circ}\text{C}$) of the capping lid top, and (c) the depth (mb) of the lid from surface to top for 1400 and 1700 UTC, 20 May, 1979.

6.4.4 Capping Lid Intensity

The above conditions fit rather well the average 'lid' conditions given by Fawbush and Miller (1954) and Beebe and Bates (1955), among others. Still, the analyses presented thus far fail to quantify the intensity of the capping lid. To estimate lid intensity, the 201400 and 201700 capping lids were examined in terms of the temperature difference between the base and top of the stable capping layers. A spatial analysis of this field is presented in Figure 6.22, superimposed with the low-level jet streaks at 201200 (from Figure 6.2). Values for this analysis were manually extracted from tephigrams, and included any layer between 900 and 700 mb where the temperature lapse rate was less than dry adiabatic (as defined by potential temperature lines), allowing negative values to appear in the analysis. The least stable regions of

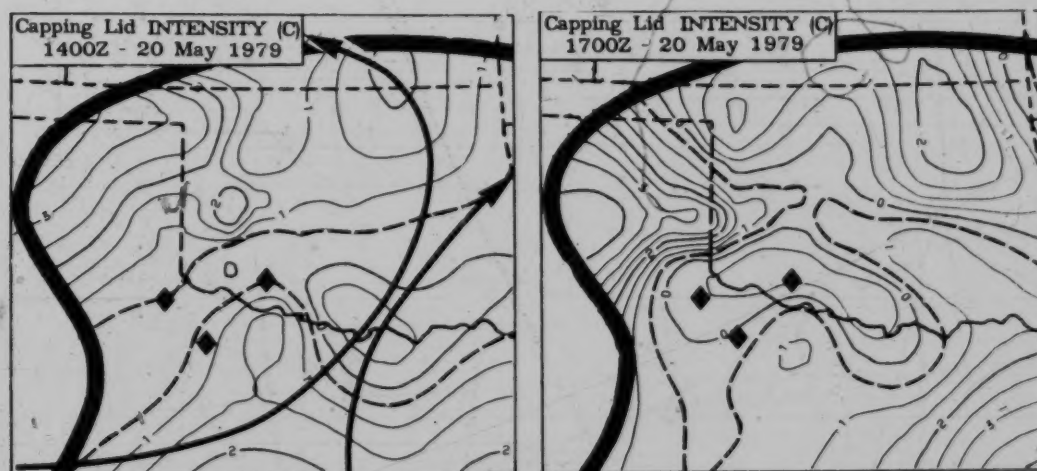


FIGURE 6.22: Capping lid intensity ($^{\circ}\text{C}$) as defined by temperature difference from the lid top to the base at 1400 and 1700 UTC, and low-level jet streaks (arrows) at 1200 UTC, 20 May, 1979. Broken lines indicate regions where the capping lid is weakest. The heavy solid line is the boundary of the lid (from Figure 6.20).

the lid (temperature inversions $< 0.5^{\circ}\text{C}$), and therefore the regions most susceptible to the occurrence of convection, have been outlined by a heavy broken line. The regional extent of the lid (from Figure 6.20) is indicated by the heavy solid line at both times. Values outside this boundary have no real meaning¹⁶.

At 201400, the least stable portion of the lid (values $< +0.5^{\circ}\text{C}$) lies over southern Oklahoma and southwestward to west Texas (Figure 6.22a). The lid is non-existent at Midland. The most stable portion of the lid lies across northern Oklahoma, the Texas Panhandle, and across northeastern Texas with prominent bulges northward, particularly through Wichita Falls (SPS, near southwest Oklahoma). These bulges, representing stable capping lid regions, correspond to similar patterns of high surface mixing ratio in Figure 6.16. It would seem that the jet streaks, which are not resolved in time here, played a key role, first in channeling the moisture northward into moist tongues, and secondly in accentuating the underrunning described by Carlson et al. (1983), where the moist air flowed out from under the more stable capping lid into southwestern Oklahoma.

6.4.5 Breakdown of the Capping Lid

There are two main mechanisms for the breakdown of a capping lid: adiabatic ascent at the lid level, and surface heating. Both mechanisms

¹⁶ Contours outside the heavy line in Figure 6.22, where there is no identifiable capping lid, are misrepresented. This resulted from the objective analysis procedure used, which extrapolated beyond where there was data. In retrospect, it was recognized that potential temperature differences would provide a better means to define lid intensity.

are illustrated in the series of low-level soundings for the Wichita Falls (SPS) site (Figure 6.23). One observes the gradual lifting and cooling of the lid, as well as adiabatic heating of the boundary layer. Some 100 km east-northeast from SPS, Healdton (HEA) exhibited an almost identical capping inversion at 201100 (Figure 6.24). By 201700, some 1.5 hours before the first radar echoes, the regions of most stable capping lid (values $>+0.5^{\circ}\text{C}$) had formed a conspicuous line running northward through eastern Texas, swinging northwestward, then westward across central Oklahoma and the Texas Panhandle (Figure 6.22b). The stable bulge persisted through Wichita Falls, but it had weakened.

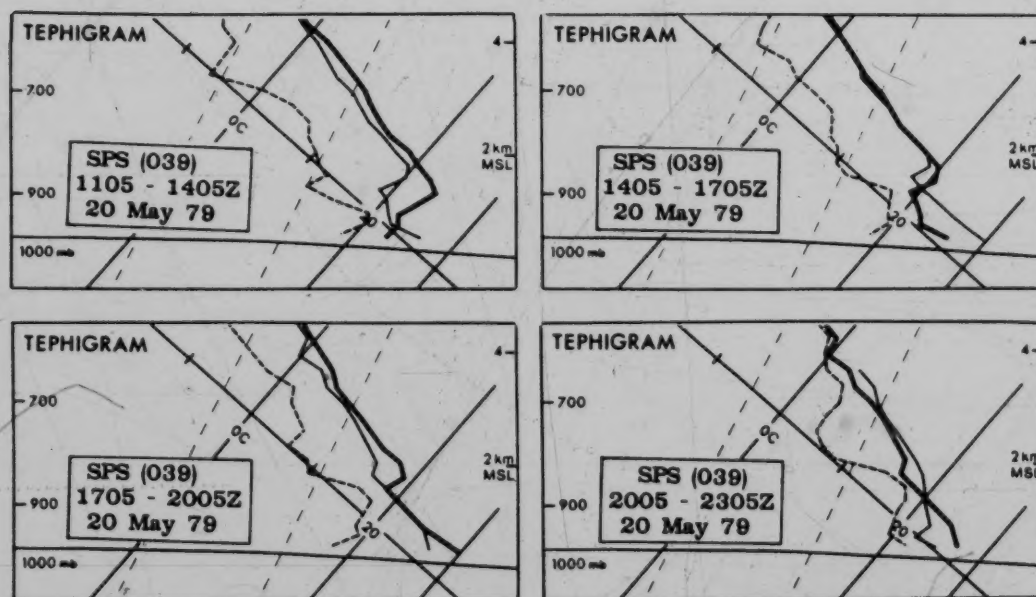


FIGURE 6.23 Low-level portion of soundings for Wichita Falls (SPS), Texas, 20 May, 1979. The lighter solid lines are temperature traces three hours later.

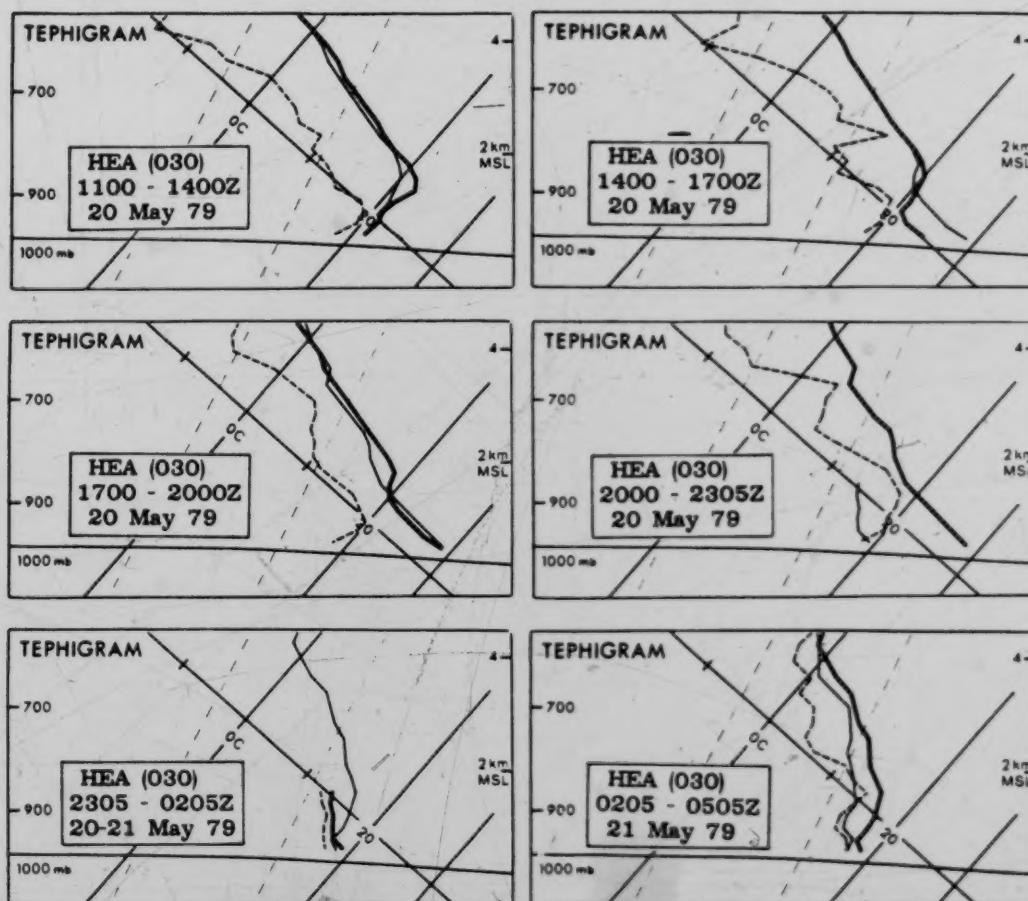


FIGURE 6.24: Low-level portion of soundings for Healdton (HEA), Oklahoma, 20 May, 1979. The lighter solid lines are temperature traces three hours later.

Over southwestern Oklahoma, where the major storm later formed, the capping lid had destabilized considerably because of ascent and surface heating over that region (Figures 5.7, 5.9). There was now significant underrunning of moist air into both southwest and southeast Oklahoma. The AVE-V storm complex initiated over the former region first for three reasons: first, because it was the region of weakest capping lid; second, because low-level ascent and moisture convergence were prevalent

there (Figures 5.6-5.9); and third, because this region experienced the most rapid increase in surface heating (Figure 6.8). A few soundings east of the storm formation area still exhibited a capping lid after 210000. In all cases, however, these lids were cooler and at a higher elevation than before.

While either adiabatic ascent or surface heating can destroy the capping inversion, in this case, however, and in most similar Alberta cases with which the author is familiar, the inversion breaks down through adiabatic cooling caused by ascent, before surface heating can significantly affect the inversion. Because of this, it rarely seems necessary to attain the convective temperature predicted from the pre-storm sounding (Figure 6.5) before convection ensues. The inability of surface heating to break down the lid in many cases was illustrated in the pre-storm sounding (with heating trend) shown in Figure 4.5c for an Alberta severe hailstorm case. This feature of the capping lid separates this type of inversion from nocturnal radiation inversions. The nocturnal inversion is generally capped closer to the surface (25-50 mb above ground), and it is usually removed by surface heating.

6.5 AVE-V FORECASTING CAPABILITY

The original impetus behind this study was our inability to predict severe convective weather with consistent accuracy. In discussing the AVE-V severe storm case in the past two chapters, various aspects of the dynamics and thermodynamics have been emphasized which, given this data set, would have been helpful in forecasting some sub-synoptic to mesoscale features of storms during the AVE-V period. For example,

detailed analyses of upper and lower jet intersections could alert the forecaster to severe storm threats over southern Oklahoma and the Midland (west Texas) regions 3-6 hours in advance, depending on the time required for the analysis. Computations of surface wind convergence, as given by Sikdar and Fox (1983), could have provided an excellent representation of the sub-synoptic scale shape of the AVE-V complex (Figure 5.6f) at least four hours prior to radar echoes of convection.

The simple adiabatic technique developed here for computing vertical velocity fields, suggests its use as a mesoscale tool for predicting the dynamics which trigger storms such as AVE-V, up to six hours in advance, along with their general shape, size, and location. However, these techniques fall short of a forecast of actual storm intensity. Missing here is a mesoscale model which might take the computed vertical motion field to predict clouds and precipitation potential. An advanced model which could further simulate the effects of local topography, surface insolation, and soil moisture effects, may make accurate 3-6 hour mesoscale convective forecasts a reality.

In the meantime, the author has developed an objective, statistical technique which provides objective forecasts of convective complexes. The technique is built around the Synoptic Index of Convection, first developed to forecast maximum hail size (the Sc10) in Alberta (Strong, 1979), then later revised to predict the location of initiation, size, motion, and maximum intensity of convective complexes (the Sc4 - Strong and Wilson, 1981, 1983b,c). The development of the Sc4, and guidelines for its operational application are reviewed in Appendix III, while several examples of its use were introduced briefly in Section 1.3. While initial development of the Sc4 was commenced several years prior

to the current research, all refinements to the Sc4, and most of its operational evolution, were developed in parallel with the author's preparatory background studies of scale interactions (Strong, 1983).

It is thus appropriate to investigate how successfully the Synoptic Index would have predicted the storms of this day. In particular, it is useful to test how accurately it could forecast the initiation regions, sizes, motions, and the maximum intensity of these storms. This discussion is based on the Sc4 guidelines outlined in Appendix III.

Some exceptions to the Sc4 definitions and guidelines are necessary here. The Sc4 is comprised of four predictors: briefly, these entail two instability predictors called the EGDEX (due to Checkwitch, 1971), and the SPLI (a simple surface lifted index), both of which are computed from observed surface data and a forecast of late afternoon 500 mb temperature, and two synoptic variables consisting of the forecast 24-hour changes in 700 mb temperature and 1000-500 mb thickness. Operationally, the synoptic variables are obtained from the latest CMC spectral prognosis, received via national teletype circuits. Errors in the 24-hour changes are relatively small, and therefore rarely cause significant errors in the Sc4 value. EGDEX computations, however, can presently only be computed for Alberta and immediately-adjacent regions, because these involve diurnal changes in the the surface pseudo-adiabatic wet bulb potential temperature (Θ -W), which were derived from 30-year climatological means for Alberta.

Consequently, the Synoptic Index, as used for this Oklahoma-Texas application, and the Colorado example given in Section 1.3, was computed using only three variables (an Sc3). Since the EGDEX is statistically

the most important variable (see Table A3.1, Appendix III), its omission more than compensates for any slight advantage in using observed rather than forecast upper air data here.

For Alberta operations, Sc4 (or Sc3) forecasts are most often based on surface data at 1500 UTC, although the procedure can be carried out using data valid at any time. By 1500 UTC, some initial boundary layer mixing from surface heating has taken place, so that large fluctuations in moisture variables do not occur thereafter, as is often the case immediately after sunrise. The same assumption was made for convection in Oklahoma-Texas.

6.5.1 The Forecast Sc4 Field and Its Verification

Following the forecast procedure of Appendix III, Figure 6.25 contains, in addition to the Sc3 forecast, the 1500 UTC adiabatic equivalent potential temperature (Theta-E) analysis, along with the latest (1200 UTC) 700 and 850 mb synoptic analyses that would have been available at that time. In the procedure used for Alberta, the 700 mb level was chosen for estimates of Sc4 geostrophic advection, since it is the level closest to average convective cloud bases, while the 850 mb level is used for boundary layer moisture advection. The same levels were used in this analysis, so as not to bias the procedure. Since the predictand, CDC, has upper and lower bounds of $+5/-3$ (see Appendix III), Sc3 values outside this range are truncated to $+5$ or -3 in the figures, which follow, for ease of reference.

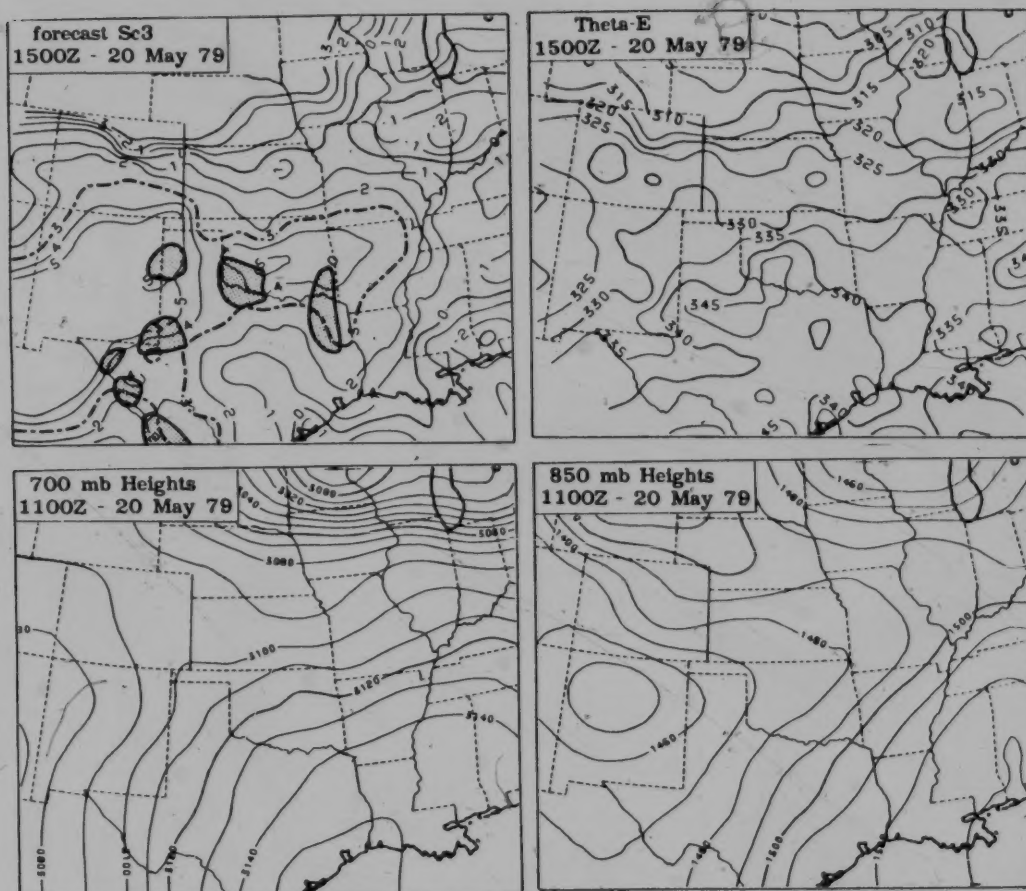


FIGURE 6.25: (a) 1500 UTC forecast Sc3, (b) 1500 UTC Theta-E analysis, and 1200 UTC (c) 700 and (d) 850 mb height analyses, 20 May, 1979. The "severe storm threat" region is highlighted by the broken (+3) contour, while predicted regions of convective complex initiation (stippled regions) are determined by the procedure outlined in Appendix III.

We note here that the major forecast severe storm "threat region" ($Sc3 > +3$) includes the western half of Arkansas, most of Oklahoma, the northwestern third of Texas, and virtually all of New Mexico. Within this synoptic scale region, however, using the most recent 700 mb analysis, one can define seven mesoscale regions of "negative Sc3 advection" where severe storms are most likely to form, as indicated by

the hatched regions of Figure 6.25a¹⁷. Once formed, these storms tend to propagate with the mean flow towards or along the nearest line of Sc3 maximum.

The major storms did indeed form in the approximate locations indicated (see radar composites of Figures 5.7-5.16), with the Oklahoma storms moving east-northeast, and the Texas storms northeast. Other storms which are not indicated on Figure 6.25 did not achieve severe intensity, with the exception of an echo which briefly reached 17.3 km at 202130 over the centre of New Mexico (Figure 5.11). This is not so surprising, considering it occurred where the forecast Sc3 had the highest values on the chart, but it would not be specifically predicted using this technique because there was no 'negative' Sc3 advection.

The proviso then, is that one should always treat large areas of forecast high Sc3 ($> +5$) with caution, even when there is no indication of negative advection. In cases where the Sc4 (or Sc3) is high without negative advection, Alberta experience suggests a forecast of widespread, but less-organized, weak to moderate thunderstorm complexes. This is essentially what occurred over New Mexico in this instance. Based on the negative advection and the absolute values of Sc3 involved, the most severe storms would have been forecast to occur over southwest Oklahoma and the Midland region of Texas. We have already seen that this was indeed the case (Figures 5.7-5.16).

¹⁷ Routine automatic computations of Sc4, Theta-E, and their advection have been implemented since these analyses were completed. During 1984 operations, the advection model correctly predicted convective regions more than 90% of the time, while the forecast Sc4 was correct on maximum intensity over the AHP operations region more than 75% of the time.

Once the initial prediction using the Sc4 and its advection has been made, the Theta-E analysis (Figure 6.25b) and its advection as implied by the 850 mb analysis (Figure 6.25d), would be used to choose the southwest Oklahoma and west Texas storm cells, once formed, as the most significant storms, since these had the highest thermodynamic potential (Theta-E). The high Theta-E values over Oklahoma also leave the strong impression that, of the two or three main AVE-V storm cells (see radar composite for 201935, Figure 5.10), the one over extreme southwest Oklahoma, once formed, had the highest potential for severe development.

The Sc4 pattern can also be used to predict 3-6 hour motions of large convective complexes such as AVE-V. These complexes tend to propagate in the general direction of the mean upper winds, but towards a line (or centre) of maximum Sc4. For trend (1-3 hour) predictions of the smaller mesoscale features, motion should be towards the highest values of Theta-E. In this case, this was eastward for the AVE-V storm and northeastward for the Midland storm. By the time new mesoscale storm cells have formed 3-4 hours later, one has more recent data available to update the trend forecasts.

The diagnostic Sc3 and Theta-E analyses at 202100 and the "hindcast" convective complexes in Figure 6.26, are an indication of the degree of consistency in the technique being used. For example, while Sc3 values had increased over eastern Oklahoma, negative advection was clearly stronger over the southwestern portion of the state, where the storm had attained its greatest intensity (Figure 5.13).

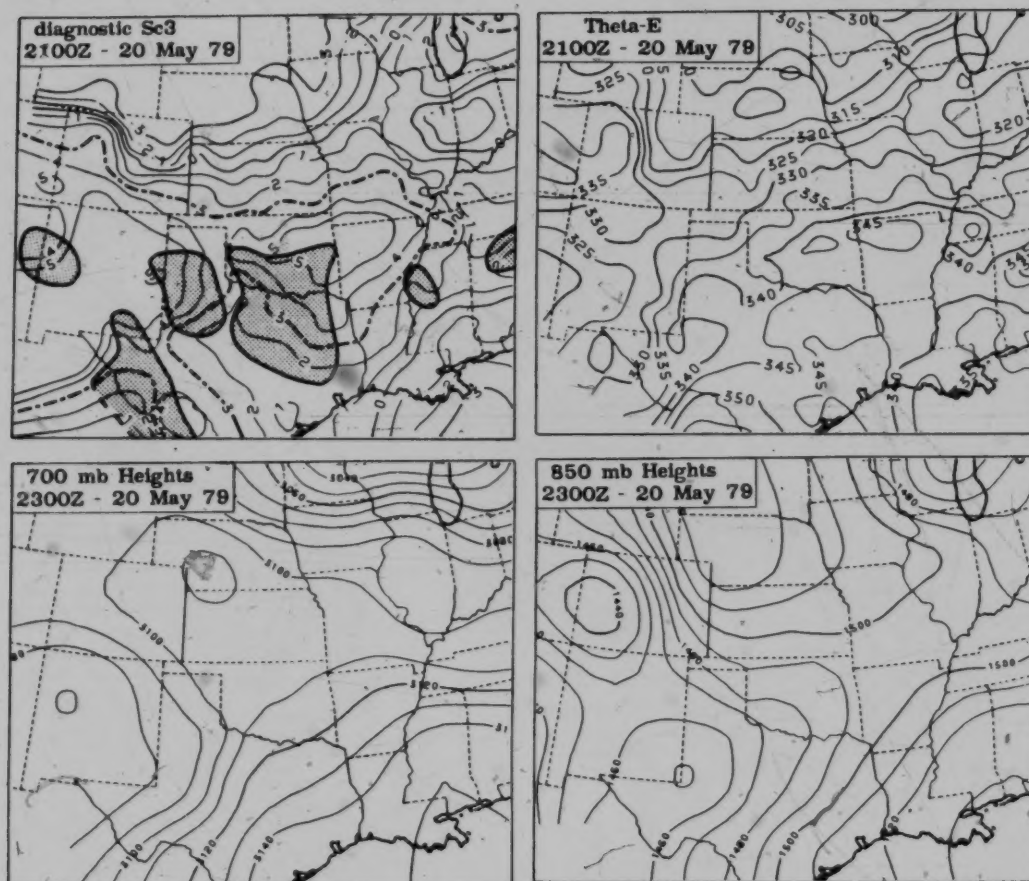


FIGURE 6.26: Forecast verification: 2100 UTC diagnostic (a) Sc3 and (b) Theta-E analysis, and 2400 UTC (c) 700 and (d) 850 mb height analyses, 20 May, 1979.

6.5.2 Comparison of Sc3 and Theta-E Fields with Vertical Motion

Figure 6.27 is a composite of the 1500 UTC forecast Sc3 ($= +3$) contour, severe and moderate convective complexes (as depicted by radar composites), reports of tornadoes or funnel clouds, and isohyets (in mm) of total rainfall for the 24-hour period ending at 211200. The Sc4, rather than its advection, is used in this composite summary, since the Sc4 was designed to predict maximum convective intensity for the full

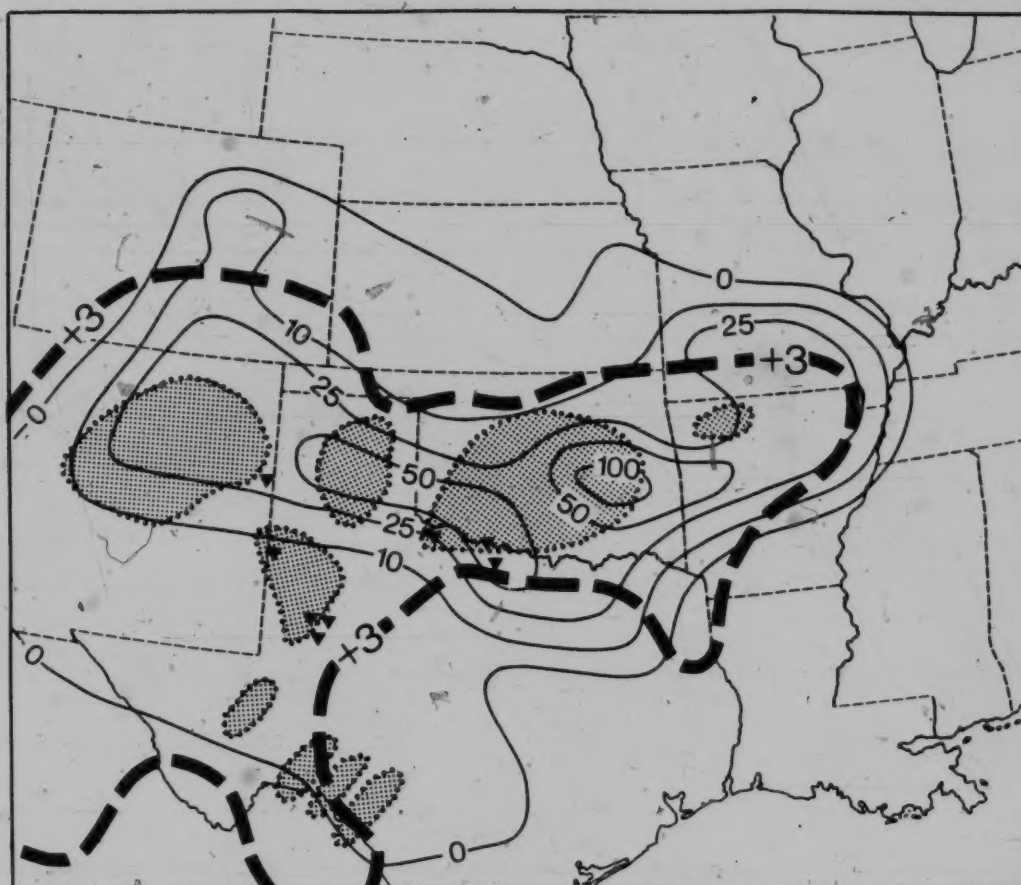


FIGURE 6.27: Forecast verification: Composite summary of forecast Sc3 contour ($= +3$), observed severe and moderate convective complexes (stippled) near times of maturity, observed tornadoes and funnel clouds (inverted triangles), and 24-hour rainfall isohyets (mm), 20-21 May, 1979.

21-hour period, 1500-1200 UTC, without regard to when storms might occur. Qualitatively, the patterns of Figure 5.27 compare favourably in terms of locations, patterns, and intensity, with the vertical motion fields of the early and mature storm period (Figures 5.9-5.13), as well as with the analysis of surface convergence at 201400 (Figure 5.6f) computed by Sikdar and Fox (1983).

6.6 SUMMARY OF THE AVE-V CASE

Two major synoptic vortices affected eastern North America during the AVE-V period. The structure of the upper jets suggested synoptic scale interactions and energy transfer between the two systems. The northern vortex, centred over Hudson Bay, appeared to be receiving kinetic energy from the southern system centred over northern Mexico. This is supported to some degree by Printy and Fuelberg (1984), who found a net transport of energy out of both the southern synoptic domain and out of the mesoscale convective region. This system-to-system synoptic scale interaction is part of the total energy budget, which warrants consideration in future multi-scale diagnostic studies involving convective storms.

The upper and lower jet structures suggested that the conceptual model of Beebe and Bates (1955) can aid in the correct identification of potential severe storm regions, even though the jets were not well-defined in this instance. This identification can be made before consideration is given to thermodynamic instability and surface heating. The fact that these kinematics have been shown to be valid, in itself suggests dynamic interactions between synoptic processes and convection, since one is using only synoptic scale data to infer processes at a smaller scale. The inference here is of a downscale transfer of energy to the pre-storm environment. Some of these effects were then examined in the mesoscale data.

6.6.1 Mesoscale Findings

Good qualitative agreement was found between synoptic scale analyses of the modified adiabatic vertical motion fields and synoptic cloud structures, and between mesoscale vertical motion fields and the observed convective complexes throughout their life cycle. In particular, convective storm initiation was preceded by significant low-level ascent and surface convergence, while the regions of ascent tended to lose their organized centres once convective storms were in advanced states of decay. Even after sunset, some storms continued, and in some cases, intensified where there was significant ascent and a continued supply of moisture.

Three possible sources for the mesoscale vertical circulations were investigated. The first two, low-level convergence caused by early local shallow convection, and lifting due to the southward movement of the cold front, were ruled out as primary causes, although convection likely enhanced the fields once mature storms were established. Adding vertical cross-sections to the analysis, these ascent regions appear to be representative of propagating mesoscale circulation systems, which developed upstream (southwest and west) as a result of a downscale transfer of eddy kinetic energy from the synoptic scale to the observed scale of motion. This was supported by the coincidental northeastward motion of the ascent centres and the convective complexes, and during the pre-storm period, by the motion of mid-level clouds related to ascent.

The energy expended by these regions of ascent in lifting and cooling the environment, both before and during the storm, might be looked upon as large-scale energy made available to the convective storm. Its contribution to overall storm energy is proportional to the area of environmental cooling on an adiabatic chart.

Some data shortcomings and a conflict of computational methods deserve emphasis here. Confirmation of temporal continuity of the centres of ascent, and also of source regions, was restricted by limitations in the domain size of the mesoscale network; and in the sounding time domain for the pre-storm period. Thus, this study cannot unequivocally establish the assumed relation between the synoptic and mesoscale processes, even though other features of the study provide no alternative explanation to the author.

Computed values of vertical velocity during the mature storm period, using the modified adiabatic method, conflict with values obtained from the kinematic method. Kinematic values of ascent in the literature of $20\text{--}80\text{ }\mu\text{b s}^{-1}$ across distances of 150 km or more, as computed from environmental winds at mid- to high-levels, have been interpreted as point values of vertical motion caused by storm-scale motion imposed on the environmental field. Such values conflict with theoretical representations of storm environment flow (e.g., Ogura and Cho, 1973), and with the few available measures of vertical motion in the immediate vicinity of storms, such as those of Fankhauser et al. (1985).

In contrast, the adiabatic method tends to reflect only the environmental component of vertical motion which the radiosonde observations are supposed to represent, and, in the AVE-V case, yields subsidence at these levels as implied by the Ogura and Cho theory. This subsidence is supported by the synoptic scale vorticity fields. The clarification of concepts of storm environment flow and the validation of vertical motion computations are important challenges for future measurement systems, possibly using the newer Doppler wind profiler measurements.

The analysis of thermodynamic stability of the AVE-V soundings concurred with similar findings by the author using Alberta data. That is, that storms do not necessarily form over locations where thermodynamic instability is maximum. Rather, they tend to form on the upwind side of these locations, where the thermodynamic gradients are strongest, and where advective changes in these instability variables are taking place. This is the case with the Synoptic Index as well.

Not surprisingly, the moisture source region for the AVE-V storms was a relatively thin layer of high mixing ratios over the Gulf of Mexico. The capping lid, a low-level jet streak, and the particular topography over and downstream from storm regions, are important factors in channeling the moisture into either a region or a line where the lid is weak, so that storms can be most easily initiated where the moisture undercuts the lid. The lower jet streaks are indirectly coupled to the upper jets through the mesoscale vertical circulations, referred to by Uccellini and Johnson (1979) as indirect circulations. The lid weakness can occur within a region of ascent associated with these vertical circulations.

The ascent initially results from eddy kinetic energy being transferred downscale from synoptic scale processes. The combination of ascent to cool the capping lid over preferred regions, and the moist air underrunning the regions of most stable capping lid, were instrumental in initiating the AVE-V severe storms. Other important factors for all storms are the local enhancement of the mesoscale regions of ascent by both topographic forcing and convergence due to surface heating. These sub-mesoscale effects are fixed local factors, which are deterministic to a large degree. Surface heating is controlled by a combination of capping lid depth, insolation, determined (locally) by cloud coverage, and surface albedo, controlled by topographic and vegetation factors.

The occurrence of the capping lid is attributed to four factors: (1) early large-scale subsidence warming, with later enhancement by (2) orographic subsidence warming from mid-level flow off the Mexican Plateau, (3) elevated mixing of hot, dry air over the Plateau, which moves out over Texas and Oklahoma, and (4) a southeasterly flow of cooler, moist boundary layer air from the Gulf of Mexico.

It was demonstrated that the major sub-synoptic scale convective complexes could be predicted (hindcast in this instance) using the Synoptic Index, without the benefit of the mesoscale upper air data. This index was developed for forecasting Alberta hailstorms. The forecast technique applies the surface temperature and moisture fields to refine the forecast to near-mesoscale resolution.

6.6.2 A Scale Interaction Summary

The available data, although lacking spatial resolution southwest of Oklahoma, suggest that several mesoscale circulation systems, with horizontal dimensions of 100-300 km, originated from the major synoptic system over northern Mexico. One of these features, identified in this study as a region of ascent at low to middle levels, was over southwest Texas at 201230, and moved northeast at $35-40 \text{ km hr}^{-1}$. Its advance provided adiabatic cooling of the lower troposphere, particularly of the capping inversion layer near 800 mb. Coupled with other convectively positive factors, such as high humidity underrunning the capping inversion over southwest Oklahoma, orographic lift, and rapid surface heating due to local surface albedo and insolation factors. A severe convective storm, the AVE-V complex, ensued.

The pre-storm cooling, and subsequent continued cooling of the environment around the storm, represents potential energy made available to the storm from the larger scale flow. The energy conversion processes involved here are not clearly understood. However, the sequence of events appears to be a downward transport of kinetic energy to lower levels upstream from the storm region. This is supported in the AVE-V case by temporal decreases in jet stream kinetic energy, especially southwest of the convective storm region, coupled with low-level increases in kinetic energy west and northwest of this region. This resulted in an increase in baroclinicity at low levels west of the storm region.

These facts are demonstrated by the intrusion of the low-level northerly jet (LJ3 in Figure 6.2), and continued cooling at low and middle levels throughout the data period (Figure 4.7). Another way to view this process, is that the increase in baroclinicity at the observed scale of motion leads to mesocyclogenesis at that scale, manifested as the mesoscale regions of ascent in this instance. The process may be similar to the scenario described by Sechrist and Dutton (1970) for the interactions between two synoptic scale systems.

A question which requires detailed attention in the future is, what are the factors which determine the spectrum of eddy dissipation, or, given specific conditions, are certain scales favoured? That is, is it possible to determine in advance what scale(s) of motion will be favoured in the eddy dissipation of kinetic energy from the synoptic scale? Ramage (1976) speaks of 'turbulence bursts', which determine the spectrum of eddy dissipation. He concludes that these bursts are basically non-deterministic.

The author remains optimistic that such is not entirely the case, and this is supported by persistent success in accurate forecasting of near-mesoscale convective events, using the simple Synoptic Index with its input of synoptic to sub-synoptic scale data. This optimism is shared by others. For example, Paine and Kaplan (1977) have presented a theory to explain how the concept of a cascade of kinetic energy, and concentration of vorticity through the hydrodynamic spectrum, could be used to numerically predict the location and intensity of tornadoes based on larger scale data sets.

A qualitative observation by the author, from his own forecasting experience in Alberta, is that mesoscale convective complexes tend to form when the larger synoptic system is in a state of decay (cyclolysis). This is so in the AVE-V case, where the surface low pressure system, initially over New Mexico, gradually lost its circulation and dropped southeastward, disappearing altogether as a separate entity by 210200 (see Appendix IVa). By the same token, the small frontal wave associated with the AVE-V complex, intensified only briefly under the influence of the convective complex, and did not have adequate resolution to be traced unambiguously back to its source.

Over Alberta, organized, severe convective complexes are not often (if ever) associated with large, intensifying synoptic systems. In such cases, convection, though prevalent near and behind the cold front from spring through fall, tends to be of smaller scale (showers and thundershowers) and less organized. The larger hailstorms experienced in Alberta usually coincide with weaker, or decaying synoptic shortwaves, which are still easily recognized. However, the author advises caution in generalizing these notions, as the observations require quantitative confirmation.

"The evolution of a severe storm system depends on interactions between motion and thermodynamic fields on scales ranging from that of the midlatitude cyclone with dimensions of thousands of kilometers, to that of submicron-diameter cloud nuclei."

D.K. Lilly, 1975

CHAPTER 7

SESAME - AVE-IV: 09-10 May, 1979

7.1 INTRODUCTION

The AVE-IV case is presented only as a confirmation test of the techniques used in this research. Because the major convective storm in this instance formed outside the mesoscale network, there is less opportunity to investigate mesoscale features of the storm, or of interactions with the synoptic scale processes. However, this case study is interesting at both the synoptic and mesoscales, since it involved such a variety of severe weather types, ranging from heavy snowfall in Colorado (091400) to tornadoes over northern Texas and Oklahoma (July and Turner, 1980). Unlike the AVE-V period, it has received fairly wide attention in the literature, even though the convective complex, for the most part, remained north of the mesoscale network.

The AVE-IV case pertains to the development of severe thunderstorms along a cold front over the Texas-Oklahoma panhandles and southern Kansas regions late in the day on 09 May, 1979. These storms, with radar echoes as high as 20 km, combined to form a major sub-synoptic

scale convective complex, as shown by the GOES visible image in Figure 7.1. This complex bordered on the northern edge of the SESAME mesoscale network throughout most of the AVE-IV data period. Figure 7.2 provides a schematic picture of the major cloud systems and flow patterns at 100000, when this storm had reached maturity.

Ogura et al. (1982) attempted to delineate triggering mechanisms in the AVE-IV pre-storm environment, and concluded that the mesoscale ascent associated with frontal lifting, was a major factor in weakening the capping inversion, thereby allowing severe storms to form rather

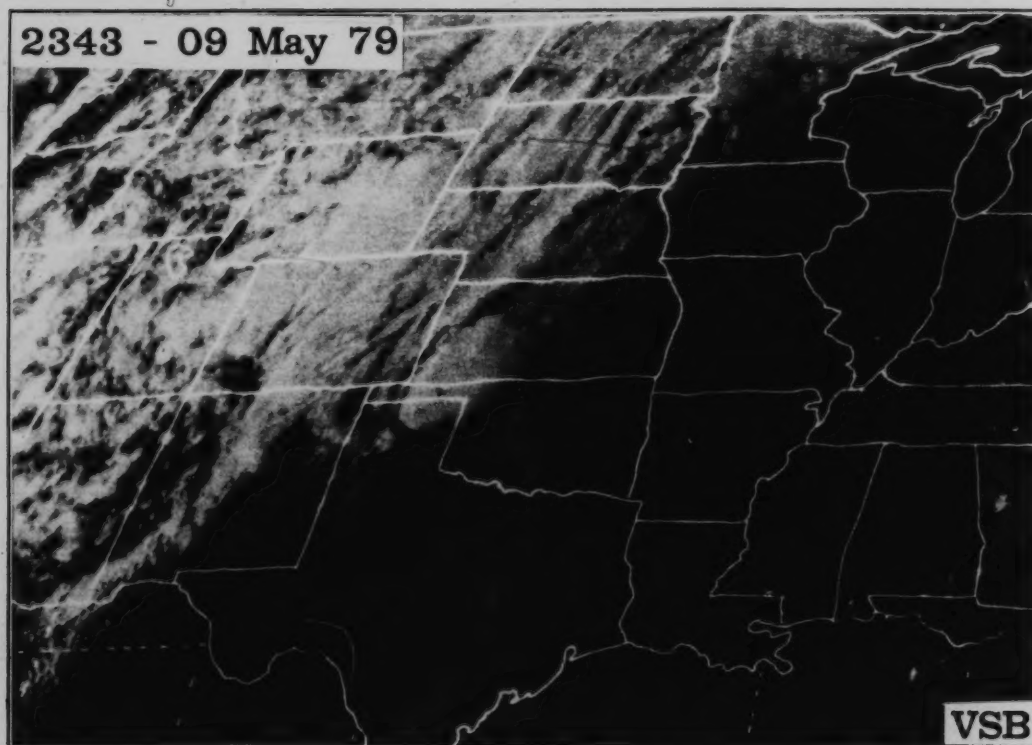


FIGURE 7.1: GOES Visible image for 2343 UTC, 09 May, 1979, showing sub-synoptic scale convective complex over Texas-Oklahoma panhandle region. (Features should be shifted slightly eastward due to a gridding error.)

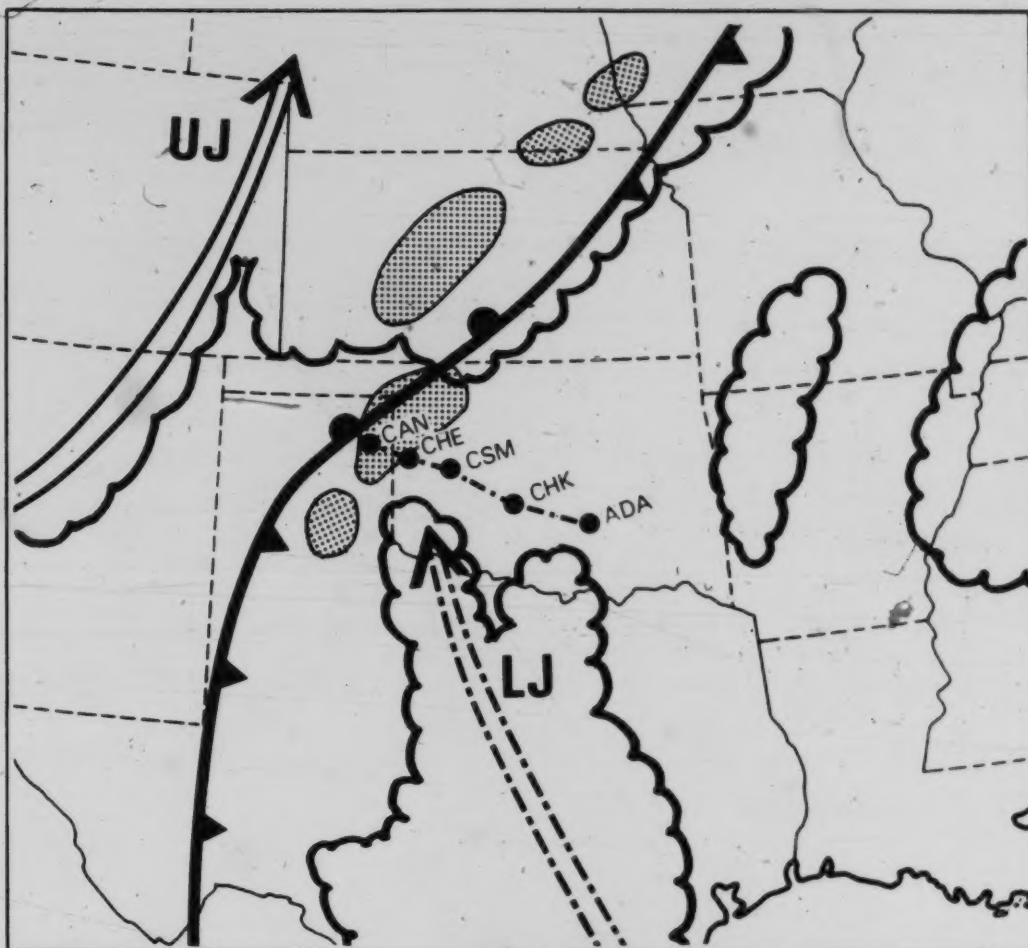


FIGURE 7.2: Schematic of cloud systems and flow patterns at 0000 UTC, 10 May, 1979. Line of mesoscale upper air stations is a reference for the soundings in Figure 7.6.

suddenly. Stobie et al. (1983) studied the severe storms which occurred earlier on the same day north of the SESAME network. These storms were associated with the same synoptic system, and are still recognizable in Figure 7.1 along the northward extension of the same cold front from eastern Nebraska to Minnesota. Stobie et al. concluded that large-amplitude, long-period gravity waves were an integral part of the

entire system of severe storms. They suggested that the gravity wave trains extracted energy from the basic state synoptic flow all along their track, through "critical level interaction". In one instance, these wave trains appeared to initiate convective cells. Carlson et al. (1983), and Lanicci and Carlson (1983) also dealt with the AVE-IV case, describing the processes of elevated mixing of hot, dry air over the Mexican Plateau, and of underrunning of moist, boundary layer air as primary causes of the conspicuous capping lid and of the subsequent storms in this case (see Figure 7.6).

Since the major convective storm formed outside the mesoscale network, a less detailed analysis is presented. The synoptic scale flow patterns and mesoscale analyses of the adiabatic values of vertical motion, as well as pertinent results from other authors, will be compared with the locations, intensities, and types of severe storms which formed late on 09 May.

7.2 SYNOPTIC SUMMARY

The broad synoptic features of AVE-IV are similar to those of AVE-V (Figures 5.1-5.2 and 7.3-7.4). Both cases involved surface low pressure centres over New Mexico which were filling while they dropped southeastward, all the while providing a southeast boundary layer flow of moist air under a capping inversion. The large area of cloud over the eastern portions of Texas and Oklahoma (Figures 7.1, 7.2) is a deck of mostly low-level stratus clouds, reflecting the boundary layer moisture feeding into the storm complex from the southeast. Both cases initially involved a quasi-stationary cold front near the mesoscale

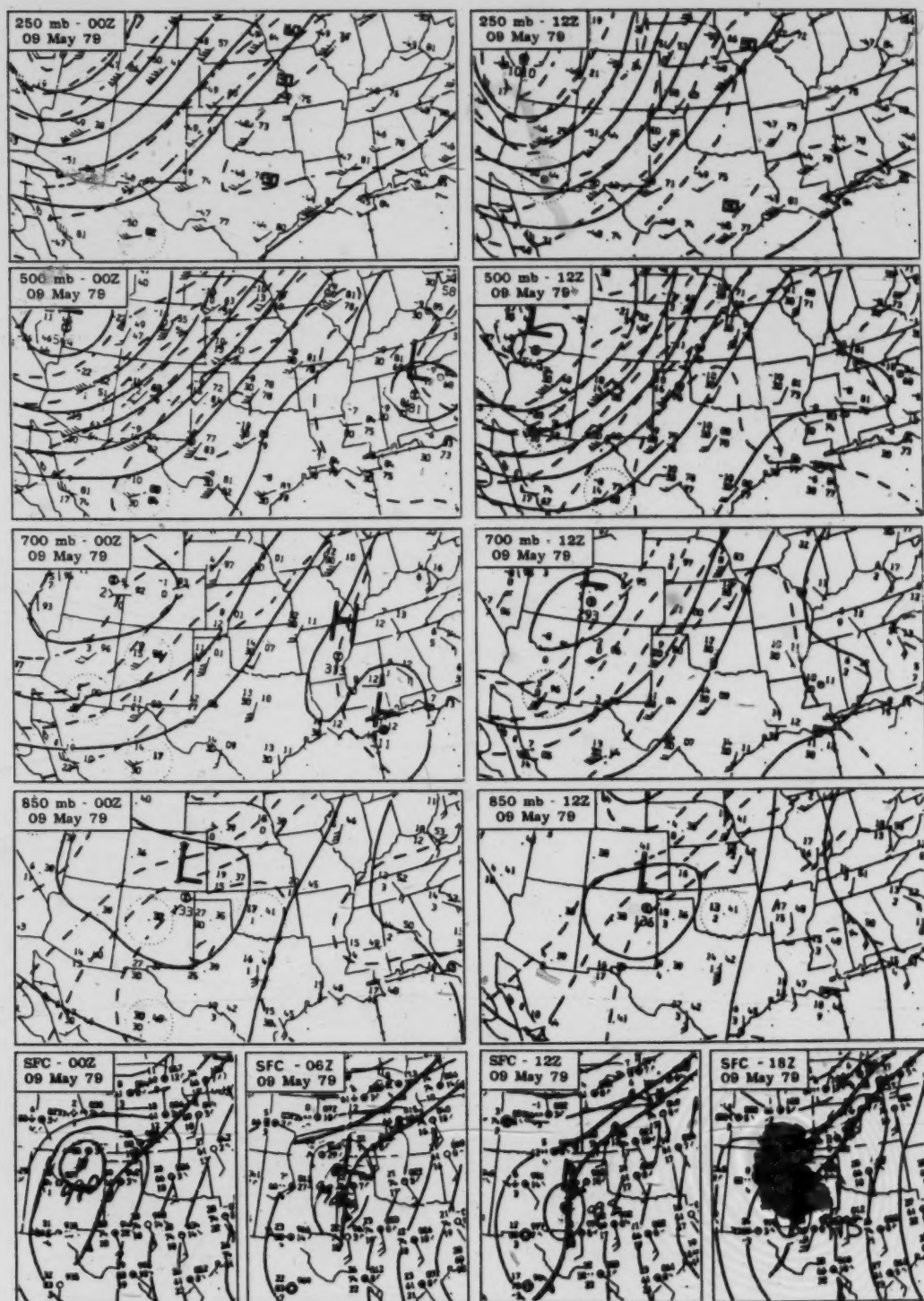


FIGURE 7.3: Synoptic analyses: surface, 850, 700, 500, and 250 mb at 0000 and 1200 UTC, 09 May, 1979.

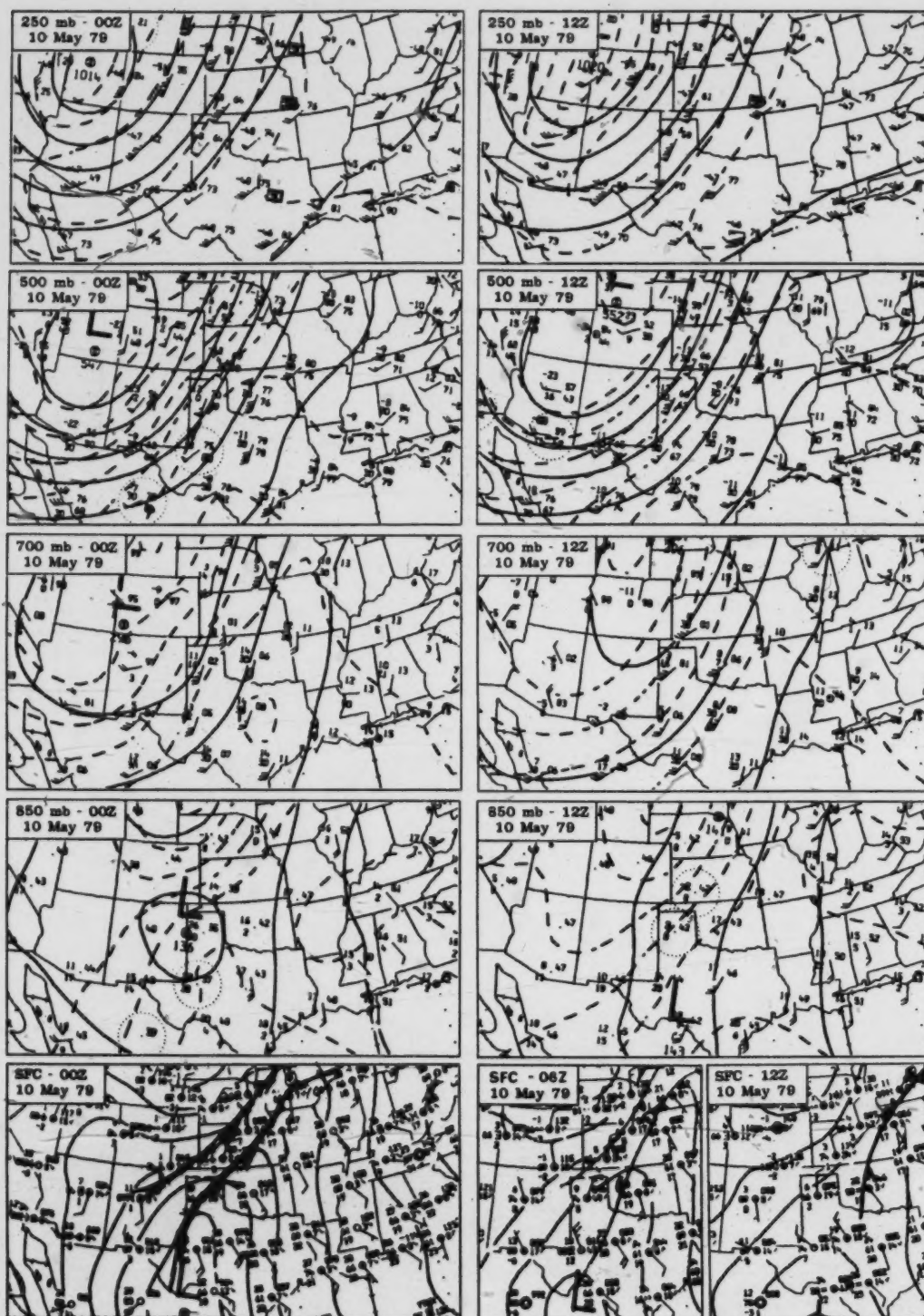


FIGURE 7.4: Synoptic analyses: surface, 850, 700, 500, and 250 mb at 0000 and 1200 UTC, 10 May, 1979.

network. The winds veered with height, with a jet stream off to the west.

However, there the similarities end. The controlling AVE-IV upper low, which included a deep trough extending southward well into Mexico, approached from the northwest (Oregon-Idaho), whereas the AVE-V low originated over Mexico. Once reaching southern Colorado (Figure 7.4), this low recurved toward the northeast. The major differences, however, were in the intensity of the upper features. A stronger baroclinic zone was evident east of the AVE-IV upper trough, at the eastern edge of which lay the surface cold front. Stronger winds, and hence higher levels of kinetic energy, were apparent at all levels, emphasized by a jet core with winds up to 150 knots (77 m s^{-1}) at 250 mb. The 500 mb height and vorticity analysis for 100000 is presented in Figure 7.5a. It suggests only very weak synoptic scale PYA over the severe storm region.

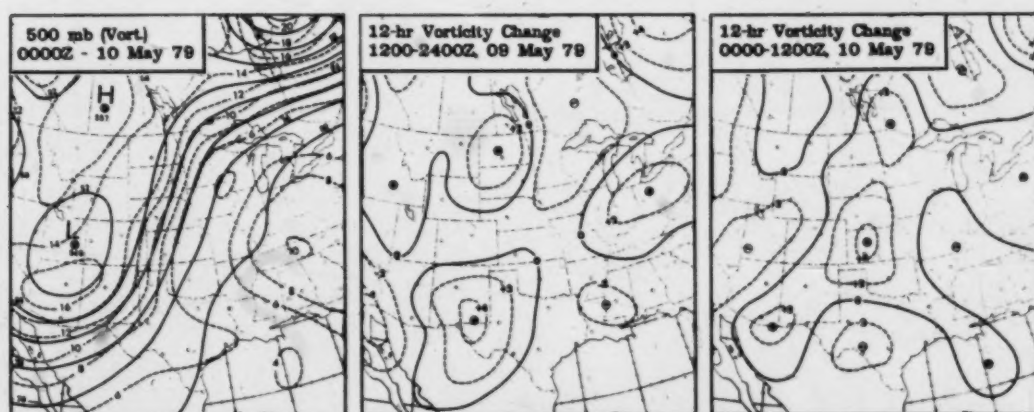


FIGURE 7.5: Synoptic analyses: (a) 500 mb height/vorticity analysis for 0000 UTC, 10 May, 1979; 12-hour vorticity changes, 09-10 May, 1979 for (b) 1200-0000 UTC, and (c) 0000-1200 UTC.

On the other hand, the 12-hour changes in vorticity (Figures 7.5b,c) show that that an area of moderate vorticity increases (changes greater than $2 \times 10^{-5} \text{ s}^{-1}$ in 12 hours) moved northeastward at $50\text{--}55 \text{ km hr}^{-1}$ from southeast New Mexico, across the severe storm region to western Kansas between 091200 and 101200. One can view these 12-hour changes as the approximate vorticity advection valid at 091800 and 100600 respectively. This PVA is representative of significant synoptic scale ascent, which, from extrapolation, would have been centred over the northern Texas Panhandle at 100000, about the time of severe storm maturity (see Figure 7.1).

The locations and motions of the centres of PVA correspond well to the surface low centre at 091800 (Figure 7.4), and subsequently, to the major storm complex and reported storm motions (July and Turner, 1980). However, since the vorticity advection analyses were derived from CMC synoptic analyses which had input data only of synoptic spatial and temporal resolution, one cannot reasonably make inferences of mesoscale storm patterns.

A few other features of the synoptic flow deserve to be pointed out here. Figures 7.3-7.4 show that the upper low, over northern Arizona at 091200, recurved from a southeasterly direction towards the northeast at least 24 hours prior to the AVE-IV storms. This change in motion had nothing to do with severe storm effects, but it may well be tied to the severe storm cause. The associated surface low, as in the AVE-V case, was in a state of decay throughout this period, despite the PVA support aloft.

In spite of apparent warm advection over the mesoscale network during the severe storm period (after 092100), the mid-levels (850-500 mb) cooled significantly over the storm region after 100000. Over Oklahoma City and Amarillo, for example, the 700 mb level cooled by 6°C and 8°C respectively between 100000 and 101200 (Figure 7.4). This can be attributed partly to cold air advection as the cold front slipped eastward, but the PVA suggests that adiabatic cooling due to ascent was just as important a factor.

While no detailed wind analysis is presented here, July and Turner (1980) noted that the jet stream winds upstream from the severe storm over New Mexico diminished in intensity during the storm period, while an increase in jet core winds is perceptible in the synoptic analysis downstream from the storm. From similar trends in the AVE-V wind fields, it was speculated that a downscale transfer of kinetic energy to the upwind flank of the storm was taking place, with a positive feedback to the synoptic scale flow downstream. Perhaps a similar process was happening here.

7.3 SOUNDINGS AND THE CAPPING INVERSION

The single most noteworthy feature of the AVE-IV data set is the intense capping inversion, -evident on all soundings southeast of the storm formation region. Figure 7.6 provides a cross-section of the capping lid region from the early morning (091100) and mid-afternoon (092000) soundings, for the line of stations indicated in Figure 7.2. This line runs normal to the southwesterly mid-level flow, from the northwest (back) edge of the capping lid region, near CAN where the

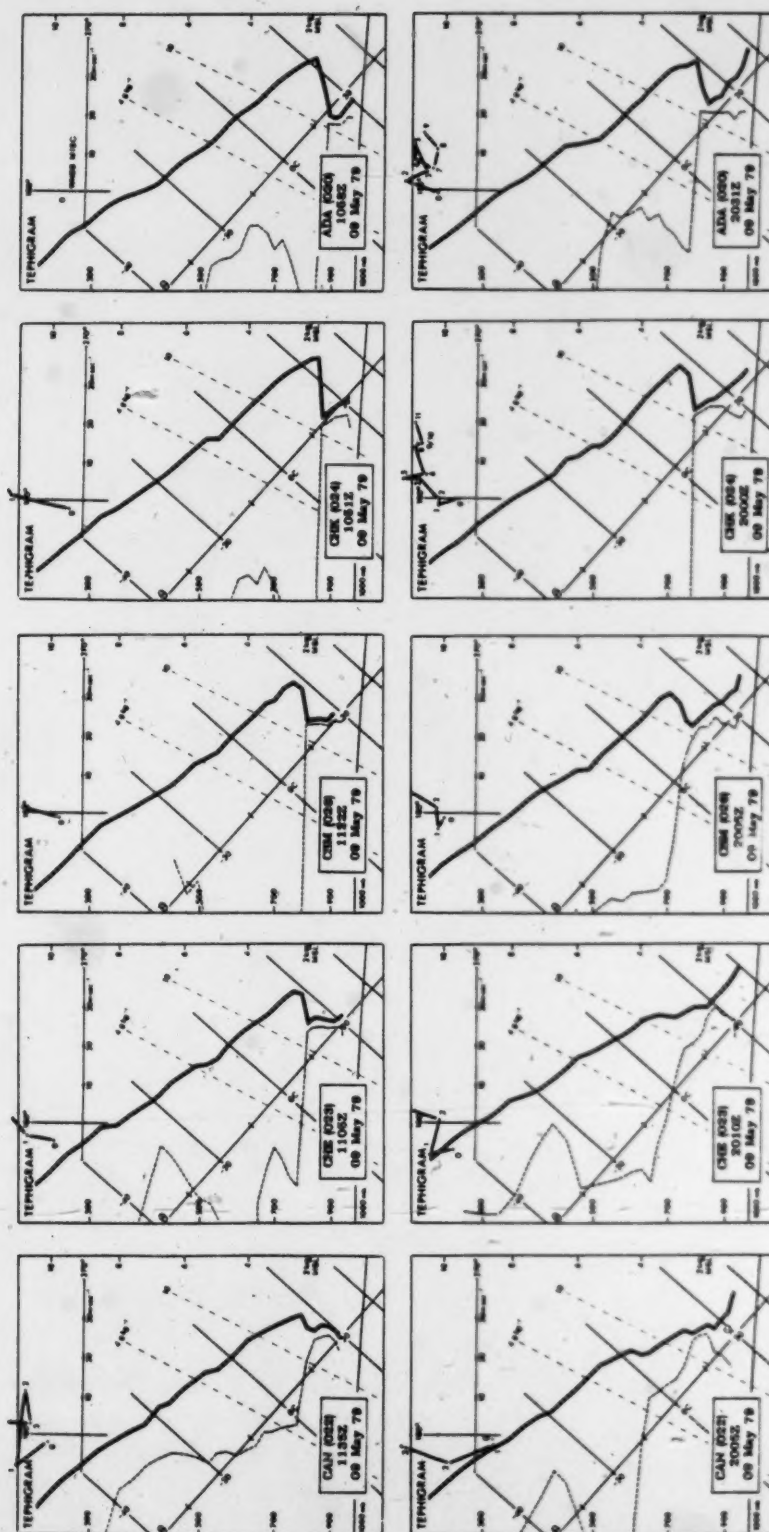


FIGURE 7.6: Sounding profiles exhibiting capping inversion strength at Canadian (CAN), Cheyenne (CHE), Clinton-Sherman (CSM), Chickasa (CHK), and Ada (ADA), at 1100 and 2000 UTC, 09 May, 1979.

storm was initiated, to ADA, deep into the stable region. On the morning (091100) soundings, the inversion was capped around 775 mb near the eventual storm initiation region (CAN), but it was stronger and 50 mb lower over the eastern end of the mesoscale network (820 mb at ADA). By mid-afternoon, when convection had just commenced over the Oklahoma and Texas Panhandles, ~~only a trace~~ of the lid remained over the storm region (at CAN and CHE), because of pre-storm ascent and adiabatic cooling. At the same time, the inversion had lifted (cooled) to the 750-700 mb layer over the eastern sites, where weaker ascent was probably occurring. By this time, a well-mixed moist layer was clearly evident, especially on the western-most soundings. Carlson et al. (1983) provide a more detailed analysis of the AVE-IV capping inversion.

7.4 VERTICAL VELOCITY FIELDS AND STORM DEVELOPMENT

A time series of the modified adiabatic vertical motion fields at 800 and 700 mb is presented in Figures 7.7-7.9. The synoptic domain is shown, although the analyses are at three-hour intervals (sub-synoptic temporal resolution), and they include data from the SESAME mesoscale network. These patterns will be compared qualitatively with the synoptic analyses (Figures 7.3-7.5), GOES satellite cloud patterns (Figures 7.1, 7.10), mesoscale details of radar composite summaries (Figure 7.11), and additional mesoscale patterns described by July and Turner (1980), Ogura et al. (1982), and Carlson et al. (1983).

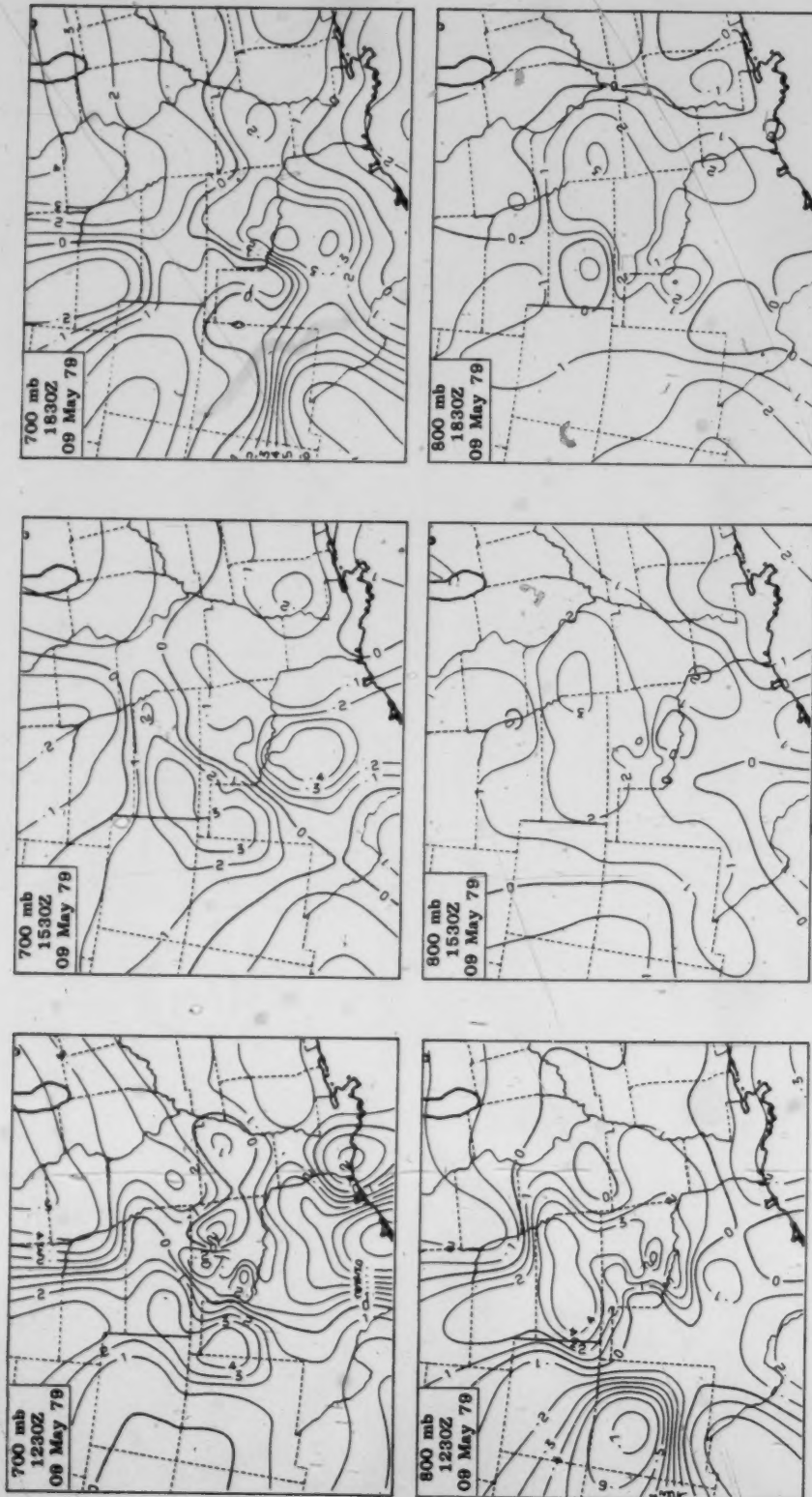


FIGURE 7.7: 700 and 800 mb vertical velocity patterns during the pre-storm period, 09 May, 1979.

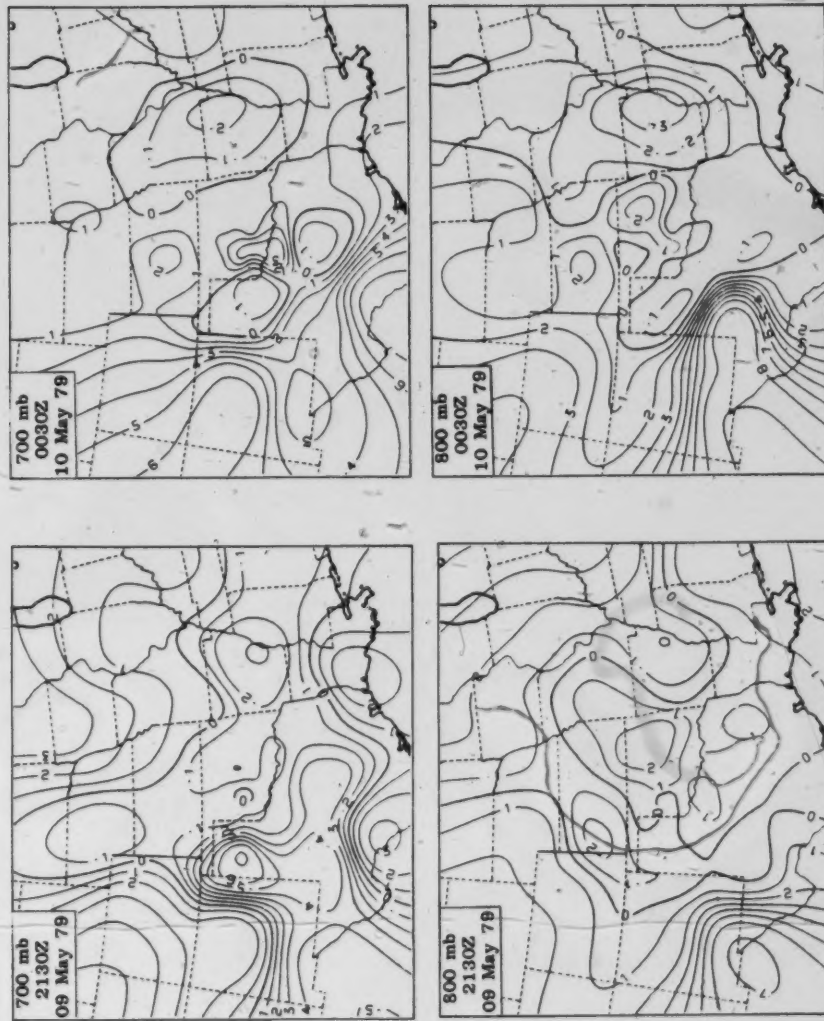


FIGURE 7.8: 700 and 800 mb vertical velocity (cm s^{-1}) patterns during the developing storm period, 09-10 May, 1979.

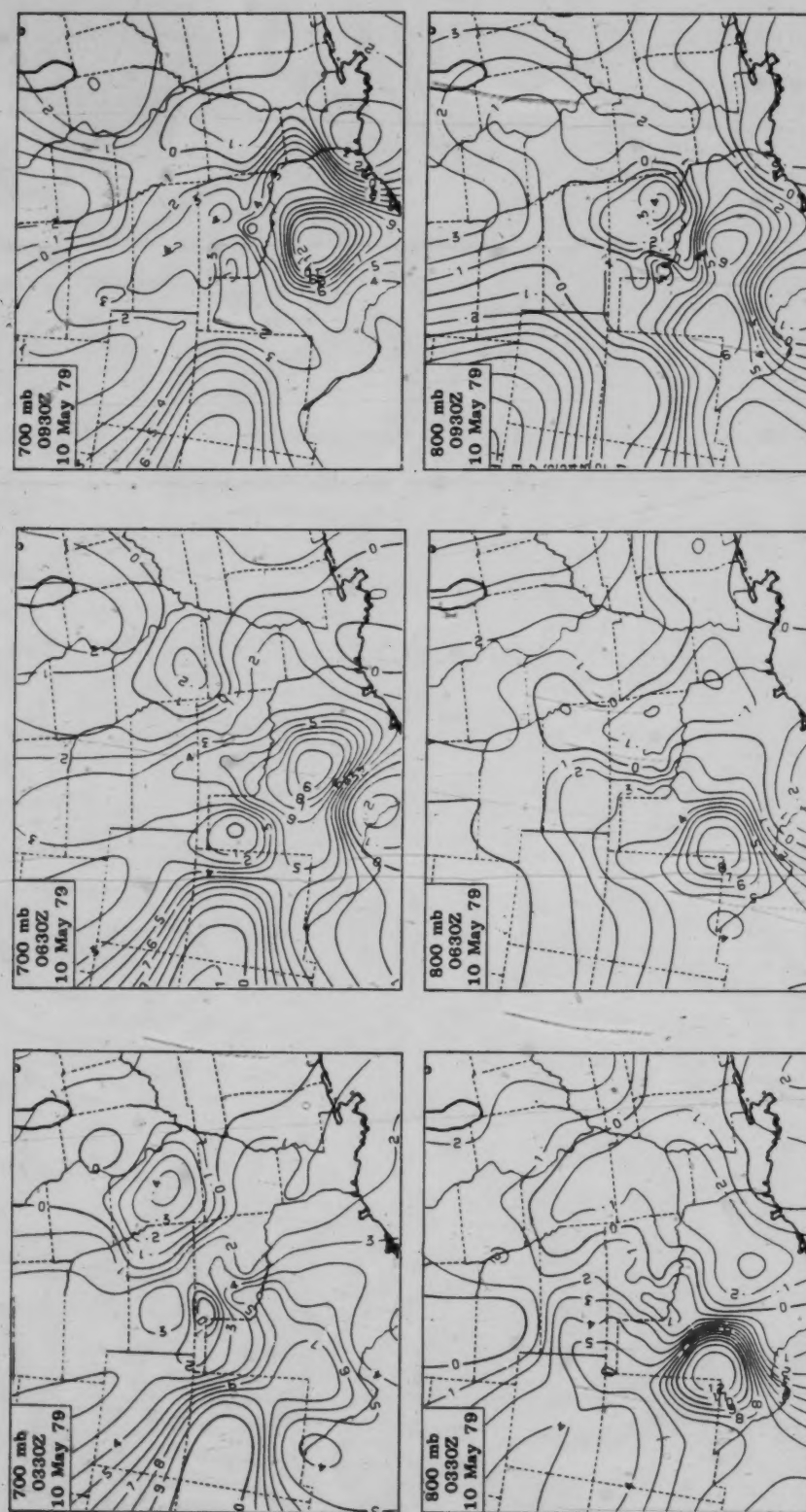


FIGURE 7.9: 700 and 800 mb vertical velocity patterns during the mature storm period, 10 May, 1979.

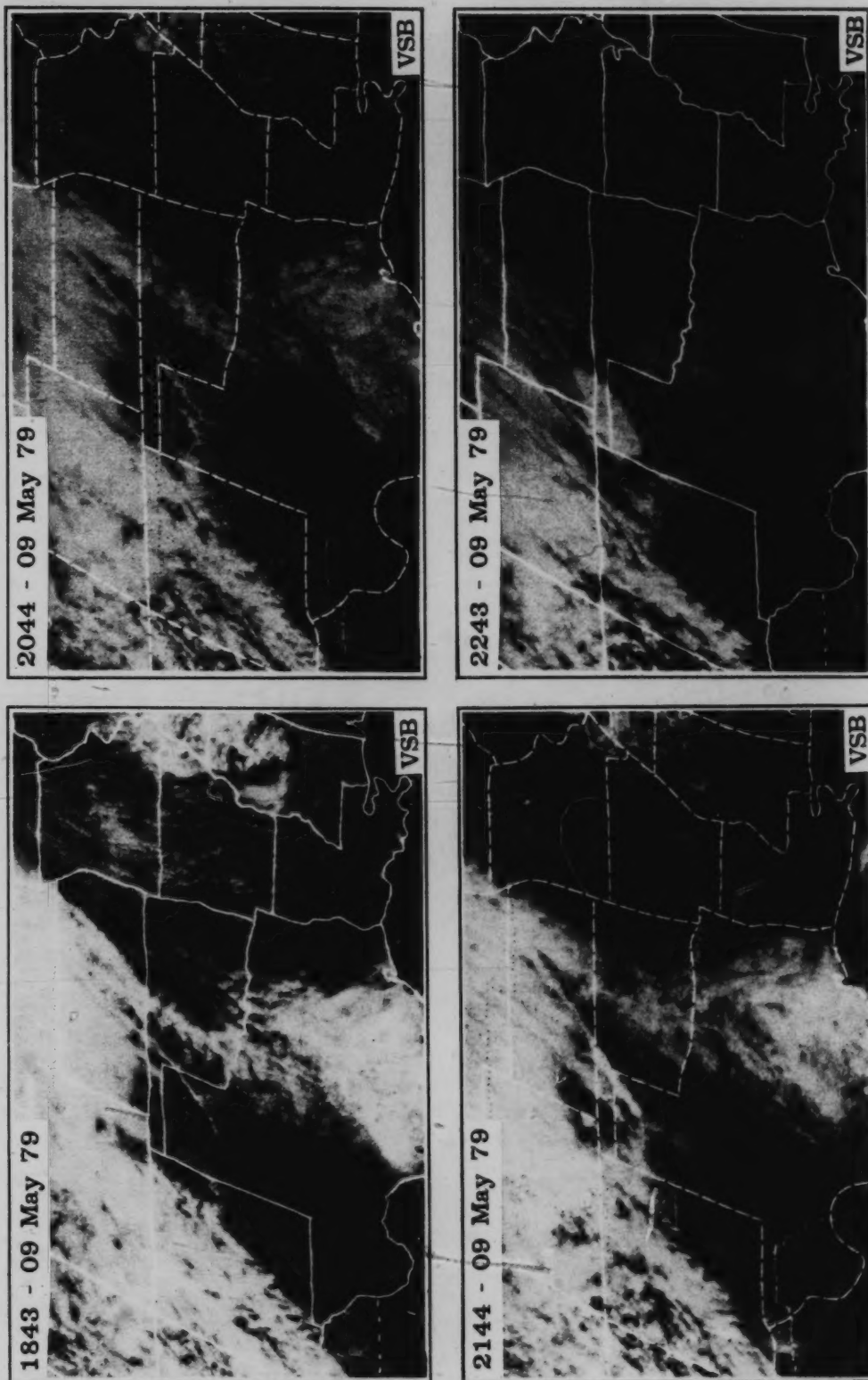


FIGURE 7.10: GOES visible images for the pre-storm (1843), early storm (2044 and 2144), and mature storm (2243) periods, 09 May, 1979. (Features should be shifted slightly eastward at 2243 due to a gridding error; see Figure 7.11b.)

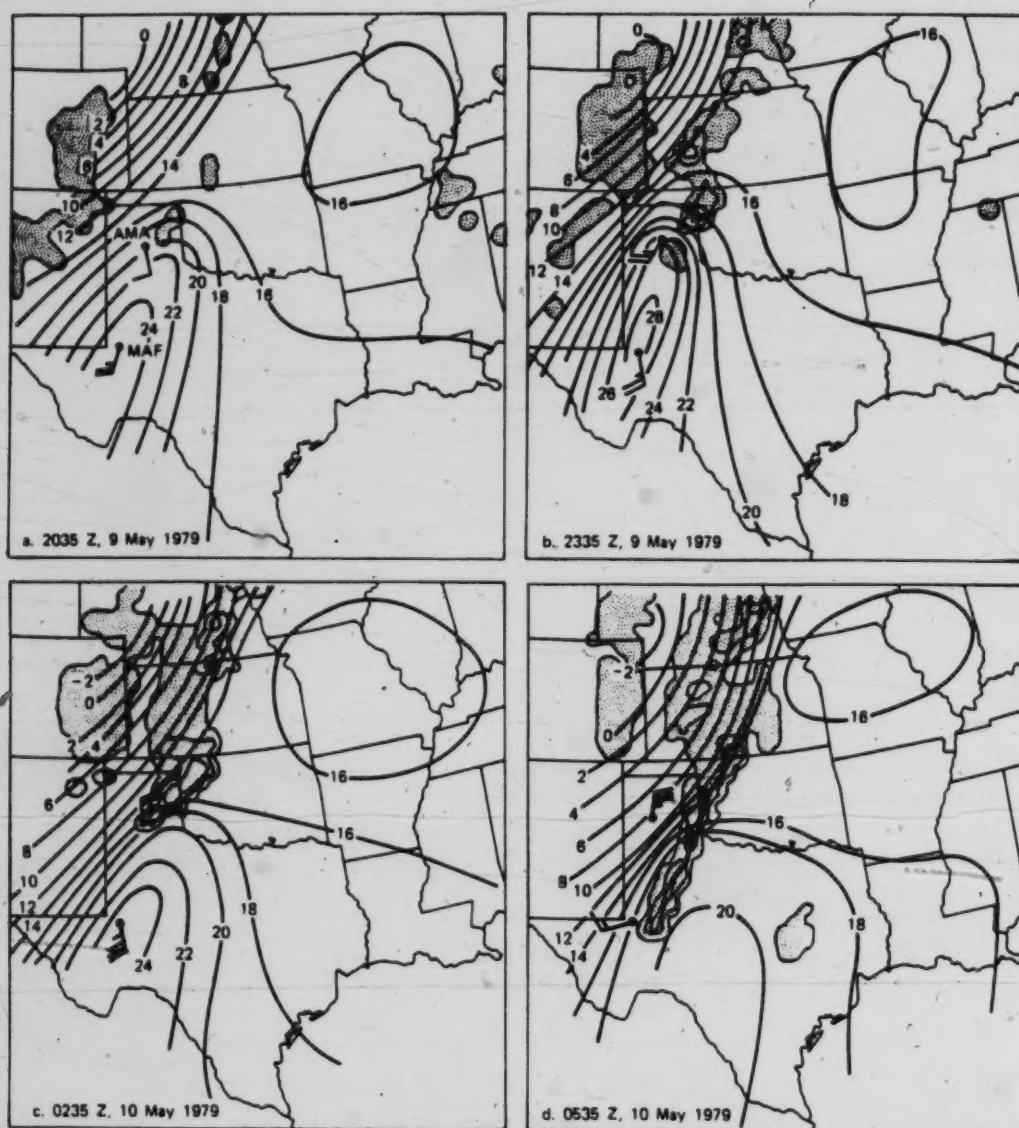


FIGURE 7.11: Radar composite summaries, isotherms ($^{\circ}\text{C}$), and 850 mb winds, illustrating the storm formation and rapid southward propagation along the cold front from 2035 - 0535, 09-10 May, 1979 (reproduced from Ogura et al, 1982).

7.4.1 The Pre-Storm Period

There are a number of important factors concerning the mesoscale region of 700 mb ascent over the Panhandle regions between 091230 and 091830 (Figure 7.7). First of all, it can be said with some confidence that this ascent was dynamically related to the decaying synoptic scale surface low and the frontal wave over that region, as well as to the region of surface convergence computed by Ogura et al. (1982). In particular, their pattern of surface convergence at 091400 matches the 700 mb centres of ascent over the Texas Panhandle and western Kansas at 091230 - 091530. Similar correspondence is found between the surface convergence and ascent over eastern Oklahoma, Arkansas, and south through eastern Texas. The low continued to fill and dropped southward out of the region by the time of storm maturity, while the frontal wave persisted in the vicinity of the main storm.

Second, the ascent was largely responsible for the gradual removal of the inversion over the Panhandle region, allowing the severe storm to form.

Third, the southwest to northeast line of storms which eventually formed through this region, was coincident with the orientation and motion of the front and the line of maximum ascent throughout the AVE-IV period. This line moved slowly eastward during the period.

Fourth, within the spatial resolution of the upper air data, one can say that other centres of ascent were oriented along the front at a wavelength of 300-400 km. Ogura et al. (1982) deduced a similar horizontal scale for the regions of surface convergence along this AVE-IV front at 091400.

The resolution of upper air data is too coarse to accurately resolve the motion of individual centres of ascent. However, while the line of maximum ascent moves eastward with the front and the squall line, the centres themselves do appear to move northeast, along with individual thunderstorm cells as depicted by radar echoes (July and Turner, 1980),

Low-level ascent also lifted the low-level moisture to form the stratiform clouds which persisted over the Panhandle regions below the inversion. Over eastern Texas and western Oklahoma, boundary layer moisture was being channeled northward by a 20-35 knot jet streak (July and Turner, 1980) underneath the strong capping inversion, the latter based between 800 and 750 mb. Low-level stratiform clouds and high humidity associated with this jet, and trapped by the inversion throughout the AVE-IV period, are evident in the GOES images of Figures 7.1 and 7.10.

By 091800, several hours before significant convection, the surface low over southeastern New Mexico was rapidly filling and dropping southward (Figure 7.3), with 700 mb subsidence across most of New Mexico and Texas (Figure 7.7). The increasing vorticity (091200 - 100000) over this region in the synoptic scale analysis (Figure 7.5b), would appear to contradict this subsidence, with a maximum increase right over the low centre. However, low-level (800 mb) mesoscale ascent was moving into the region from the southwest by 091830, and was significant at both 800 and 700 mb by 092130. It was not possible to resolve these mesoscale features with the the synoptic scale vorticity change analysis, which is a graphical difference of patterns 12 hours apart.

7.4.2 The Storm Development Period

For the most part, the convective storm, and the mesoscale perturbations assumed to be affecting this storm, passed by the mesoscale network to the northwest. Therefore, while the mesoscale patterns of ascent at 092130 (Figure 7.8) appear to be weakest during the period of storm formation and early growth, the true intensity of the ascent associated with the storm is unknown. The 700 mb patterns were not affected by errors due to diabatic heating, since the mesoscale network soundings (Figure 7.6) were basically unchanged by surface heating at 700 mb, due to the strong capping inversion. Conceivably, however, the 800 mb adiabatic computations of ascent may be underestimated because of errors due to diabatic heating.

Ascent values exceeding 8 cm s^{-1} were appearing by 100030 near Midland, west Texas, as the upper synoptic trough approached from the west. The line of maximum ascent had moved into western Oklahoma, with mesoscale centres of ascent over both Oklahoma and Kansas. However, vertical motion values from at least three key sites close to the storm complex at that time, AMA, CAN, and GAG, were all missing from 092130 on, because of missing sounding data. Because one missing sounding negates two time profiles of vertical motion using the adiabatic technique, it was not possible to accurately resolve the mesoscale characteristics of vertical motion near the main storm after 092130. Positive values of vertical motion over western Kansas (using Dodge City - DDC profiles), which increased at 100330, suggest a centre of ascent further south than shown at 100030, possibly over the Texas-Oklahoma panhandle regions where the main storm was located.

Subsidence persisted over eastern Oklahoma until 100030, and this appeared to maintain and even intensify the capping lid over the eastern sites (Figure 7.6). Ogura et al. (1982) also felt that subsidence was an important factor in the existence of the capping inversion. This conflicts with Carlson et al. (1983), who stated that "large-scale subsidence was not a factor in the development or sustenance of these lids" (capping inversions), either for this AVE-IV case or others which they had analyzed. Mid-level ascent had moved eastward across most of Oklahoma by 100030, and while the capping lid was too strong to remove with the resultant adiabatic cooling, it did lift from 800 mb to 700 mb by late in the period (not shown).

7.4.3 The Mature Storm Period and Storm Propagation

The radar composites provided by July and Turner (1980), indicate the formation of a line of severe storms after 092335, extending north-northeastward from the eastern Texas Panhandle. Individual storms moved northeastward along this line at $40\text{--}50\text{ km hr}^{-1}$, the same motion as the centres of vorticity change and mesoscale centres of ascent. The storm line progressed eastward at less than 10 km hr^{-1} , which concurs with those of the surface front and the line of maximum ascent. New storms formed so rapidly on the southern end of the line, that it had extended down into southern Texas by 100635. Storms were most intense between 100100 and 100300, when radar echoes to 20 km were reported over northern Oklahoma.

Synoptic scale vorticity increased aloft locally (Figure 7.5) as the upper trough continued eastward toward the storm region (Figures 7.3, 7.4). Strong ascent at both 800 and 700 mb also moved across central Texas from the west after 100330 (Figure 7.9), with values exceeding 12 cm s^{-1} . The radar echo tops from storms associated with this ascent over south-central Texas exceeded 16 km from 100500 through the end of the AVE-IV data period (July and Turner, 1980). Thunderstorms also occurred over northeastern Texas after 100900, but the strongest ascent remained south of Oklahoma and the mesoscale network for the most part.

Weaker ascent, and occasionally small areas of subsidence, persisted over Oklahoma, especially the eastern portion, at low levels. A number of facts support the weaker ascent over this area. The GOES images and radar summaries (July and Turner, 1980) show that eastern Oklahoma remained relatively cloud-free, and certainly storm-free at least until 101200. The capping inversion remained strong over the eastern radiosonde sites, although it cooled and was lifted due to weak ascent. Furthermore, only a small increase in synoptic vorticity occurred over eastern Oklahoma between 100000 and 101200 (Figure 7.5c).

7.4.4 Causative Factors and Source Regions

A mesoscale baroclinic wave moving northeastward from the low pressure centre and along the surface front, may have been responsible for the observed vertical motion patterns and the ensuing convective storm. However, the horizontal movement of the front itself was not at all the main cause of the vertical motion features, since, during the

24-hour period 090600 to 100600, the front moved southeastward and laterally across the storm region at less than 10 km hr^{-1} .

The storm remained primarily northwest of the mesoscale network. This fact, plus missing data from Amarillo and sites over the western portion of the mesoscale network - CAN (2300-0800 UTC), CSM (0500 UTC), GAG (2300-0500 UTC) - make it difficult to prove that the centres of ascent retained the northeasterly motion of the convective storms.

However, based on the data available, this did appear to be the case. One can argue a strong case that the source of the mesoscale perturbations in these vertical velocity fields appears to be the main synoptic system to the west and southwest. This is supported by evidence from Carlson et al. (1983). They deduced, from the temporal continuity of the surface pressure falls associated with the AVE-IV convective storms, that the falls originated from travelling upper-level disturbances. Furthermore, the continued nighttime southward propagation of the storms along the line, against the flow aloft and in the total absence of surface heating, is further evidence that the larger scale dynamics were instrumental in the initiation and propagation of the storms. The existence of the mesoscale regions of ascent simply confirms the synoptic scale evidence.

7.5 SUMMARY OF AVE-IV RESULTS

The AVE-IV case involved severe, tornado-producing thunderstorms which formed along a cold front, from southwestern Texas northward through western Kansas. The storms formed west of the mesoscale network

late in the day. The evidence presented here, supported by results from Ogura et al. (1982) and Carlson et al. (1983), shows that mesoscale ascent associated with travelling disturbances aloft, and detected maintenance of convective storms throughout the night. The mesoscale regions of ascent weakened the conspicuously strong capping lid, finally allowing convection to ensue late in the day at locations near the frontal zone, where boundary layer moisture was underrunning the inversion from the southeast.

The largest and most severe storm formed over the panhandle region (Figure 7.1), and within the scale of resolution of data, was coincident with both the pre-existing main frontal wave (Figure 7.2) and the main centre of ascent (Figure 7.7). At the same time, both the southwest to northeast orientation and the movement of the line of maximum ascent, corresponded to the orientation and the movement of both the line of convective storms and the cold front over the panhandle region.

While the patterns of ascent in this case corresponded with broad features of the cloud systems and the convective complexes, the degree of correspondence was not as clear as in the AVE-V case, for two reasons. The most obvious reason for this was the lack of mesoscale data resolution over the storm region. The storms formed west of the mesoscale network to start with, and important sounding data was missing during the period when the storms were near and over the western portion of the network. A second reason is that there is somewhat more uncertainty in the AVE-IV computed values of vertical velocity, since temperature advection, a neglected term in the simplified adiabatic method adopted, was more important in this case, as evidenced by the intensity of the cold front.

While it can be stated that the mesoscale regions of ascent came from the general direction of the synoptic scale disturbance, and were probably spawned by the same, the lack of data resolution does not allow one to pin-point source regions for these features with certainty. The greatest drawback appears to be the limitation in domain size of the SESAME (AVE-IV through -VI) networks, even when storms passed over the region as in the AVE-V case. A larger upper air network of the same spatial and temporal data resolution, possibly elongated in the direction of the average upper air flow (southwest to northeast), would allow one to establish the temporal continuity of the mesoscale features, and to resolve source regions for the oft-neglected pre-storm dynamics.

At the very least, denser surface observing networks would help to resolve sources through computations of wind divergence and moisture fields. The AVE-IV case served to again emphasize the importance of the pre-storm environment. Once the above features can be resolved and understood in terms of the energetics of scale interactions, then a total conceptual model of convective storms might be formulated.

An interesting similarity between the AVE-IV and -V cases, worthy of further investigation, is the fact that both situations involved decaying synoptic scale systems during the formation and intensification of the mesoscale convective complexes. Both were marked by filling surface low pressure systems which dropped southward, and jet stream level winds which decreased in intensity upstream from the convective formations, suggesting a drop in kinetic energy upstream at the synoptic scale. At the same time, the mesoscale motions appeared to take over and intensify.

Similar synoptic scale signatures have been observed by the author from severe storm forecasting experience in Alberta. But during situations of synoptic development (such as during lee cyclogenesis associated with frontogenesis over Alberta), extensive stratiform cloud systems and widespread continuous precipitation events tend to occur, while convection tends to be less intense in general.

Since synoptic scale dissipation (or development) events are usually evident from synoptic scale data, this may provide important clues to impending severe weather events. It may be that these events are linked in some way to the concept of "turbulence bursts" proposed by Ramage (1976), suggesting the transfer of kinetic energy downscale to the mesoscale during frontolysis or cyclolysis. Detailed multi-scale energy budget case studies will be necessary in order to investigate this hypothesis fully.

"Random advection tends to drive vorticity variations toward smaller scales of motion, until the scales become so small that diffusion can take over. This is the core of the concept of the "enstrophy cascade" in two-dimensional turbulence."

"...Vorticity fluctuations should move toward smaller scales without changing their intensity."

"...The microstructure of sub-grid scale eddy fluxes cannot generate significant changes in the evolution of the large-scale flow field."

H. Tennekes (1978)

CHAPTER 8

SUMMARY, CONCLUSIONS, and RECOMMENDATIONS

The first two chapters of this dissertation accorded extensive discussion to scale interactions and related concepts. This was necessary because of the very broad nature of the topic, covering atmospheric mechanisms from the planetary down to small mesoscale circulations, as well as convective processes. CISK, which deals with interactions of tropical cumulus with the larger scale flow, involves similar processes, but with different emphases than for mid-latitude interactions. Therefore, beyond its definition in Section 2.7.6, CISK was not considered in this study.

Significant steps in the development of scale interaction concepts have been discussed in terms of mid-latitude convective complexes, including some speculation on the reasons for past courses of research and for the remaining arguments today. Some of these are recapitulated here, followed by a summary of the author's own diagnostic results (from Chapters 3-7), based on Alberta and Oklahoma-Texas severe storms.

8.1 OVERVIEW OF SCALE INTERACTION EVIDENCE

Diagnostic studies of large scale energetics over the last 30 years have confirmed earlier theoretical work, concerning the existence of energy interactions between the synoptic and planetary scales of motion. Lorenz (1955, 1960) extended this concept of interactions to smaller scales of motion. Without identifying specific scales, he partitioned the potential and kinetic energy budget equations into zonal and eddy (perturbation) components, and distinguished between eddy and frictional dissipation of energy. White and Saltzman (1956) alluded to possible interactions between the synoptic scale and severe thunderstorm systems. However, few scientists attempted diagnostic computations of energetics at the mesoscale, primarily because of the lack of mesoscale data, but also because of the controversial nature of extending scale interaction ideas to the mesoscale.

Kung and his associates (MK72, KT75, and TK77) were the first to apply mesoscale energetics analysis to severe storm environments, using the NSSL data set. However, they experienced difficulty in attempting to relate values of the terms in their mesoscale kinetic energy budgets, especially those of generation and dissipation, back to the synoptic scale. No attempt was made to compute a concurrent synoptic scale energy budget, and this appears to be necessary to confirm such relations. Fuelberg and Printy (1984) computed both the synoptic and mesoscale energy budgets for the SESAME AVE-V case, but they considered only upscale, or feedback effects on the synoptic flow during the mature, severe storm stage. The impact of downscale effects, especially on the pre-storm environment, remains to be investigated in detail. The purpose of this thesis was to propose a hypothesis to explain such

downscale effects, and to commence the necessary diagnostic work which reveals some of the evidence in existing data sources.

Up until the 1970s, severe storm research was concentrated on the visible storm and its internal cloud and precipitation processes. Even in 1985, those concerned with formulating conceptual models of severe thunderstorms usually see little need to consider any processes beyond the storm scale and its immediate environment, as though the storm operated in a vacuum. Meteorologists from the forecasting field can take credit for early concepts of synoptic scale influences on severe storms. Beebe and Bates (1955) were the first to combine these ideas into a rational, synoptic kinematics model of severe storms, which is still useful today. Despite its survival in the forecasting field, their model was not widely accepted by those more directly involved in severe storm research. One reason for this might be their failure to relate synoptic scale fields of ascent to observed mesoscale regions of surface convergence, which many still believe are caused solely by the severe convection itself.

Scientists continued to believe well into the 1970s that non-deterministic, sub-mesoscale processes were primarily responsible for the observed mesoscale fields of convection. But meanwhile, forecasters were developing moderately successful kinematic and statistical prediction models which contradicted this view. The Synoptic Index, which incorporates synoptic scale upper air data and sub-synoptic scale surface data to predict mesoscale storm complexes, is one example of such a model. This model, developed with Alberta statistics, has now been tested successfully from northern Alberta to southern Texas, and from eastern British Columbia to Manitoba. It lends

strong support to the concept of energy interactions, especially downscale effects between synoptic scale processes and mesoscale convection, since it uses pre-storm data at one scale to predict later events at the smaller scale. Once mesoscale convective complexes were recognized as distinct entities (Maddox, 1980), often with circulation patterns similar to those of synoptic features, the concept of scale interactions gained new interest. In fact, this concept is the main driving force behind efforts to launch the multi-scale STORM-Central project (Zipser, 1984), the most ambitious meteorological field project ever.

One of the most important considerations in support of scale interaction concepts can be attributed to the theoretical works of Starr (1948), Rossby (1949), and Lorenz (1955). Between them, it was shown that frictional dissipation alone does not account for the total energy sink for any cyclone. (In view of this theory, together with the literature review and the results reported in this study, one can, with some justification, replace the word "cyclone" above with "mesoscale convective complex" with little loss of meaning.) This means that some other energy sink is involved besides molecular friction, and the term "eddy dissipation" was invoked (Lorenz, 1955). Lorenz also suggested that the dissipation can occasionally be negative in sign, since verified in some of the diagnostic studies. This factor may account for occasional "upscale" or feedback interactions from "sub-grid" processes, but normally the term is positive where cyclonic systems or severe convective complexes are involved. According to Tennekes (1978), these sub-grid scale eddy fluxes cannot generate significant changes in the evolution of the large-scale flow field.

8.2 DIAGNOSTIC RESULTS OF THIS STUDY

Six major goals for this research were stated in Section 1.4. These included the formulation of a scale interaction hypothesis for severe thunderstorms in Alberta and elsewhere. This hypothesis, described in Section 3.2, links the combined effects of synoptic processes, topographic forcing, and radiative surface heating to the initiation and maintenance of convective weather of any intensity. The problem of relating vastly different spatial and temporal scales of motion from synoptic to cloud scale, to another intermediate scale process, seems to have restricted research in this area. Thus, the kinematic models summarized by Beebe and Bates (1955), and used by the forecasting community, received little attention in research. This scale problem is removed by invoking the concept of eddy dissipation or energy transfer of the synoptic scale processes, which lead to preferred mesoscale regions of both environmental ascent and convection.

While the hypothesis was tested with detailed mesoscale analyses from only two cases, the speculation as to scale interaction causes is also based on extensive results reported in the literature, analogous storm cases from Alberta, and prolonged field observations and forecasting experience by the author.

8.2.1 Test Results of the Scale Interaction Hypothesis

Very limited Alberta data were available for this study. However, it was possible to focus on the most recognizable thermodynamic signature of severe storms throughout the high plains, especially storms

forming in the lee of the Rockies. This feature is the capping inversion, or lid. The capping lid has been under investigation by the author for a number of years, with a series of short experiments near Limestone Mountain, whence comes the experiment name, LIMEX. This region is recognized as a frequent storm initiation region within the AHP study region. LIMEX investigations were important steps in the formulation of the scale interaction hypothesis concerning severe convective storms presented in Chapter 3.

Diagnostic results of the SESAME study show that the capping lid is a link between the effects of synoptic scale processes, and the fixed, sub-mesoscale topographic and surface radiation effects on convective storms. The formation of the capping lid is linked closely to the effects of synoptic scale subsidence warming, enhanced locally by orographic subsidence, and to moist, cool underrunning or stationary boundary layer air. In southwesterly flow situations over Alberta, the orographic component of subsidence enhances the lid intensity within a narrow region less than 80 km wide over or east of the foothills peaks (such as Limestone Mountain), which lie just east of the main Rocky Mountain barrier. Over the Texas-Oklahoma region, hot, dry air mixed over the Mexican Plateau can also play a role in the lid formation, when this elevated air is advected out over the lower plains (Carlson et al, 1983). However, this effect does not appear to be a significant factor in Alberta capping lid formations. In Alberta, the orographic component of subsidence often extends to the floor of a high valley 10-25 km wide between the main mountain barrier and the foothills peaks, helping to create a dryline front, or mesoscale baroclinic zone over the foothills at the western edge of the capping lid region. Boundary layer moisture

becomes trapped over the foothills underneath the lid. There it can accumulate, increasing latent instability at the same time. This occurs in two ways: first, there are local diurnal increases in moisture because of evapotranspiration and soil evaporation with daytime heating; second, an approaching shortwave, following behind the large scale subsidence, often results in an easterly component of flow beneath the lid, advecting moisture against the foothills near the western edge of the lid region.

Weaknesses in the capping lid, which allow early convection, normally occur first over the foothills mountain ridges, where the combined effects of mesoscale regions of ascent, the links to synoptic scale energy transfer, and enhanced by topographic forcing, cause adiabatic cooling of the lower and middle layers of the troposphere. A low-level jet streak can intensify the process, by channeling boundary layer moisture beneath the lid into the weak lid region, culminating in a sudden release of convective instability.

Convective clouds tend to be organized according to the pre-storm pattern of low-level ascent and surface convergence. The size distribution of the regions of ascent depends on the relative importance of the synoptically induced vertical motion field and that caused by topographic forcing. This can vary greatly, depending on the situation, and ultimately the pattern determines the location of initiation, the size, and the shape of convective complexes. When small-scale topographic forcing is dominant, for example, thunderstorms are believed to be small, and are often single local cells which sometimes do not move off the foothills. This type can occur with or without the influence of a capping lid. When a mesoscale region of ascent is

dominant, storms may commence the same way under the influence of enhanced orographic lift, but are apt to grow into larger complexes of a scale comparable to the region of ascent. When synoptic scale vertical motion dominates, such as during rapid frontogenesis, convective storms tend to be small and less severe again, because they lack the mesoscale organization. An exception to this is along an active cold front of the developing synoptic system, where a severe thunderstorm squall line can form.

The SESAME data were used to demonstrate that weaknesses in the capping lid allow the earliest convection, and determine where the storms initiate. In the AVE-V and -IV cases, the lid did weaken where ascent was present. The ascent lifts the lid, and causes adiabatic cooling of it and adjacent layers of air, thus increasing instability to the point where free convection occurs. The ascent is also related to surface convergence of moisture beneath the stable lid region, the moisture underrunning the lid into the unstable (weak lid) region. The evidence presented for these SESAME test cases, and for similar severe storm circumstances in Alberta, suggests strongly that the mesoscale ascent is caused by energy transfer from the synoptic scale, enhanced by topographic forcing and differential surface radiation. These effects combine as a shortwave trough approaches. The trough may lead to lee cyclogenesis, accompanied by an easterly, upslope component of boundary layer moisture flux over the foothills on the north side of the surface low development. The eastern slopes of the foothills are normally heated more rapidly than the plains regions during the morning hours, a factor (anabatic wind) which probably enhances the easterly boundary layer flow.

It is important to note that in both SESAME cases studied here, extra-storm environmental ascent was present only in the lower troposphere, generally below 500 mb, during the pre-storm to mature storm stages. Deep tropospheric ascent did not arrive in these cases until much later, by which time the storms were starting to decay. If this is a general rule (and this cannot be ascertained in this study), then one requires forecast output of vertical velocity at low and middle levels from an operational numerical model. Traditional numerical output of, for example, vorticity at 500 mb, will not always reflect low-level ascent, and perhaps the analysis and forecast of 700 mb vorticity would be more suitable.

The mesoscale regions of ascent are, in effect, mesoscale vertical circulations, which in the most intense cases, also develop well-defined horizontal circulations. This happens, for example, at the mature stage of convective complexes, when a secondary low centre appears on a surface isobaric analysis, such as the small frontal low associated with the AVE-V complex between 202000 and 202300 (Appendix IVa).

Low-level jet streaks, coupled to upper jets and the larger scale flow through these mesoscale vertical circulations, are an important mechanism in the scale interaction process. The lower jet tends to intensify in response to the mesoscale ascent, topographic forcing, especially along river channels, a mesoscale baroclinic zone which can form near the back edge of the capping lid, differential heating of foothill slopes, and moisture channeling. These factors were related qualitatively but not quantitatively in the SESAME analyses. However, it was noted in the AVE-V case, that a decrease in kinetic energy at the upper jet stream levels downstream from the storm, coincided with energy

increases at low levels on the northwest side of the storm. It was not possible to confirm these results for Alberta with the minimal data available.

While topographic forcing and the location and strength of the low-level jet streaks are important additional, local effects on storm initiation and intensity, both are determined to some extent by the presence and strength of the mesoscale ascent and associated surface convergence. The ascent and convergence enhance or initiate the easterly component of flow in the boundary layer. The ultimate location of storm initiation is therefore determined by the mesoscale regions of ascent. Both the mesoscale ascent and the low-level jet are coupled to the larger scale flow by the unknown nature of large scale energy transfer, and to smaller scale flow by the "fixed", but, in principle, known variations of the topography and differential surface heating. The latter factor is dependent mainly on surface albedo and the varying terrain slope, with the eastern slopes being heated more rapidly during morning, thereby enhancing the surface convergence field and the low-level jet.

8.2.2 Continuity and Predictability of Mesoscale Processes

The computed regions of ascent for the SESAME data corresponded with mesoscale convective features over the mesoscale network, and with sub-synoptic features over the coarser network. Time-averaged spatial analyses of vertical motion compared favorably with synoptic features such as trough/ridge positions, vorticity advection, and major cloud regions. The temporal continuity and source of the ascent regions was

difficult to establish, mainly because of the limited domain size of the mesoscale network. However, temporal continuity was deduced qualitatively from related features, such as satellite-observed cloud elements, which originated in the larger synoptic scale system.

Using a time series of vertical cross-sections of vertical motion averaged over the mesoscale network, ascent was shown to be present below 700 mb more than six hours prior to first radar echoes of convection, and organized surface convergence at least four hours before radar echoes. During AVE-V, the average values of low-level ascent were $3-6 \text{ cm s}^{-1}$ over western Oklahoma. This was sufficient to have raised the lower portions of the airmass 0.5 to 1.5 km in the six hours preceding storm initiation. The importance of this ascent was demonstrated by how quickly the storm complex took on the general (horizontal) shape of the sub-synoptic scale region of surface convergence and low-level ascent of several hours before. Thus, the spatial and temporal continuity of these mesoscale convective features which preceded the storms, and their apparent synoptic source, suggests that such storms are deterministic. The success in forecasting the storm occurrences from other data sources, such as synoptic kinematics or by applying the Sc3 procedures, simply reaffirms this.

The spectrum of energy transfer (eddy dissipation) from the synoptic scale to smaller scales appears to be determined by the state of development or decay of the larger system. During synoptic development (frontogenesis or cyclogenesis), the disturbance requires vast quantities of thermal energy for transport northward and upward. Perhaps this leaves less energy for eddy dissipation to scales between the synoptic and viscous. With our current knowledge of the subject,

one can only speculate in such cases, that eddy disturbances are mostly of small scale, and that no large convective complexes result. This does not rule out lines of smaller showers and thunderstorms along an active cold front. During synoptic cyclone decay, on the other hand, the disturbance must dissipate larger amounts of energy which cannot be accomplished solely through friction. In this case, intermediate sub-synoptic or mesoscale circulations result, and subsequently, convective complexes similar in scale to these circulations may develop.

It was necessary to speculate here, since there is little, if any, theoretical or diagnostic treatment of this aspect of the study in the literature. The ideas are based primarily on the author's forecasting experience. In essence, energy from the synoptic scale results in both pre-storm cooling and continued cooling of the environment around the storm. This represents energy consumed by the storm from the larger scale flow. Multi-scale diagnostic studies of both the synoptic and mesoscales may, in the future, resolve this speculation through case study and climatological analyses of energy budget studies, by relating source and sink terms between scales.

The relative importance of sub-mesoscale topographic and surface radiation influences can be established in individual case studies through mesoscale numerical models, comparing model output to observed convective structures.

8.3 ADDITION TO THE SCALE INTERACTION CONCEPTUAL MODEL

Atmospheric processes can be described in terms of the exchanges of potential and kinetic energy between the various scales of motion. Insolation is the ultimate source for all of this energy, while friction is the final sink. But these are not alone sufficient to account for the energy gains and losses required for short periods of time over smaller portions of the hemisphere.

During winter, the major heat (energy) deficit is over the large land area at high latitudes (generally north of 40°N latitude), so that synoptic scale weather systems are the primary means of maintaining the hemispheric heat balance. Large amounts of kinetic energy are a necessary result of this large scale transport.

During summer, the regions of heat deficit are generally smaller areas, so that synoptic systems tend to lose some of their importance for maintaining the hemispheric heat balance, and smaller scale circulation systems take over. Thus, the intermediate scales between the synoptic shortwaves and viscous processes gain more prominence in the energy spectrum. The evidence for this is found in examining the size spectrum of cloud systems, representing circulation systems, on January and July satellite images of the Northern Hemisphere.

When larger heat deficit areas do occur during summer, nature responds with wave growth at the synoptic scale (cyclogenesis), so that large heat transports can take place over great distances. The result is a feedback of energy to the planetary scale, accompanied by an increase in amplitude of the longwaves, and temporarily, less energy available for dissipation to the intermediate scales. Large amounts of

APE are converted to kinetic energy during such cyclogenesis. Frictional dissipation may be an adequate local energy sink during these development periods.

Once a wave has established a latitudinal energy balance, it is then necessary for it to rapidly dissipate excess kinetic energy gained during cyclogenesis. This process, called cyclolysis, cannot be accomplished immediately by friction alone. Eddy dissipation then gains prominence as a means to overcome local kinetic energy excesses, by first transferring some of this energy downscale. This results in a series of sub-synoptic and mesoscale circulation patterns (eddies), and similar scale regions of ascent and surface convergence, as demonstrated in the SESAME cases studies. These smaller eddies are also more efficient at dissipating the excess kinetic energy through friction, while at the same time, they create preferred regions for convective outbreaks. The mathematics for this downscale energy transfer theory is suggested in the work of Paine and Kaplan (1977).

Over the oceans, the depletion of kinetic energy through friction is even more difficult. In this case, even larger, sub-synoptic scale circulations should result. Hurricanes may partially fulfill this role, at least over southern latitudes, where convection is again the major visible effect. Convection continues to convert some of the sensible and latent heat energy which was depleted earlier, back to the synoptic scale budget of kinetic energy.

This general, and in some respects, speculative concept of the summer circulation, is summarized schematically in Figure 8.1. The left side of this diagram represents stages of decay of the various scales of

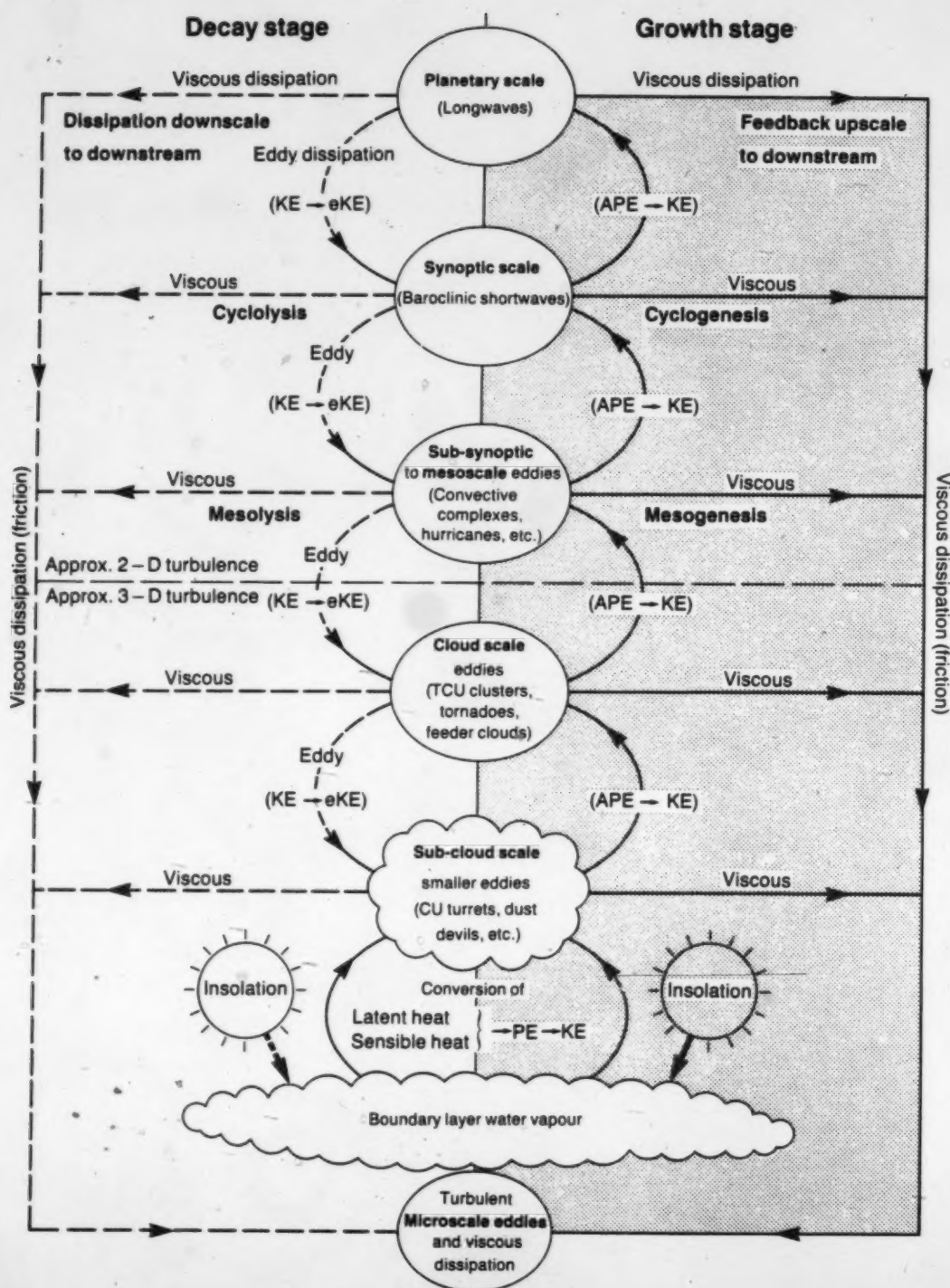


FIGURE 8.1: Conceptual model of downscale and feedback energy exchanges during decay and growth modes, planetary to viscous scales.

motion, while the right side indicates stages of development. The transfers of energy can be to more than one scale at a time, or of course, to more than one circulation system at a particular scale. The heavy arrows indicate the main perceived directions of energy transfer. For example, a developing wave at the synoptic scale will tend to supply energy to the mean (planetary) flow, which may in turn act as a source for another synoptic system downstream at a later time.

Thus, five of the six goals of this research as outlined in Section 1.4 have been achieved. The sixth goal, that of making recommendations for future field research in order to better understand these scale interactions, can now be pursued.

8.4 MESOSCALE REQUIREMENTS FOR THE FUTURE

8.4.1 Shortcomings of this Study and Future Analysis Plans

Data of many types and sources were employed in this study, and it was not possible to analyze all aspects of this data. A more complete mesoscale time series analysis of the basic upper air parameters would have proven useful; in particular, it would have been helpful to examine kinetic energy sources and sinks with more temporal detail by using the wind data. However, because of problems in the literature of interpreting dynamically-derived quantities such as kinematic vertical velocities, it was decided at the outset to concentrate on thermodynamic quantities. At the same time, adiabatic computations of vertical motion could be more easily applied later to the more limited upper air data

sets of Alberta case studies, where there are not enough mesoscale sites to perform kinematic computations of vertical motion.

The analysis of surface data was limited to the directly-measured variables (pressure, temperature, and moisture), and to forecasting applications of the Synoptic Index. Later in the diagnostic discussions, it was demonstrated how useful surface divergence computations could be, and these are included in future analysis plans. A detailed kinetic energy budget analysis has ideal applications to this type of study, but this involves a major effort by itself. An multi-scale analysis of the important energetics of the pre-storm environment, not examined by others, is a particular future goal. Attempts to map certain SESAME mesoscale features such as the capping lid, using GOES digital data, were hindered by software problems. Some GOES analysis of the Alberta case (LIMEX-80) was possible, however, including determination of the horizontal extent of the topographic subsidence effect. A detailed satellite data analysis of convective features is included in the analysis plans for the LIMEX-85 capping lid study in Alberta (Strong, 1985).

8.4.2 Recommendations

The scientific community has now arrived at a realization of the importance of scale interactions, and is preparing to study this very broad meteorological area with the multi-scale, nested grid, STORM-Central experiment, now planned for 1989. Recognizing the opportunity presented here, one would hope that Canada will participate, with an extension of the sub-synoptic scale network at least to 55°N

latitude, and a mesoscale network in Alberta.

Whether or not such a network extension is achieved, some additional suggestions would be useful for future studies. First, on experimental design logistics, the following recommendations are advanced.

1. Future large networks, designed for convective research and/or operations, should be multi-scale. In other words, the data set should be adequate for concurrent analysis at both the synoptic and mesoscales. Here, the foremost questions involve the mechanisms behind scale interactions, and the set of conditions under which energy transfers to the mesoscale are significant. This does not preclude limited mesoscale networks, designed to study a specific area or problem of scale interactions, such as the topographic effects sought out in LIMEX-80 and -85.
2. The space domain of the mesoscale network in such a multi-scale effort should be large enough to resolve the source regions of mesoscale processes. For example, it should be at least as large as one complete mesoscale vertical circulation (ascent and subsidence) of the scale studied here. The SESAME AVE-V meoscale network was approximately 350 km west-east by 300 km south-north, but was not large enough to encompass the major convective complex, even though, through sheer luck, the storm was over the network at maturity. A network of double this length, aligned with the predominant southwesterly flow aloft in severe storm situations, should be the minimum goal for the future, although it would require at least 40

upper air sites to retain the same density as AVE-V. Such a network would allow more detailed study of the life cycle of mesoscale circulations, especially of the pre-storm period, and would provide a better opportunity to identify source regions.

3. The time domain of all data sets in a multi-scale experiment should be long enough to fully investigate the pre-storm period. Data at least 18 hours prior to the convective storms being studied seem necessary to resolve scale interaction effects, and to distinguish these from the local effects of topography and of the diurnal heating cycle.

The SESAME mesoscale network unfortunately failed to adequately sample the space and time domains of mesoscale processes.

4. The upper air data must have adequate temporal and spatial resolution to resolve mesoscale processes within the constraints of sensor accuracy. The optimum rawinsonde spacing and sounding interval before instrument errors take over, appears to be in the order of 40-80 km and 1-2 hours (Fankhauser, 1969; Sackiw and Strong, 1983). The SESAME rawinsonde spacing of 75 km and sounding interval of three hours (a few of 1.5 hours), was adequate for most analysis purposes in this study. Still, smaller field experiments prior to STORM-Central should investigate the benefits of having finer data resolution. For example, the LIMEX-85 data might help in this regard, with an average site spacing of 55 km, and soundings at two-hour intervals from nine upper air sites (Strong, 1985).

5. Standard surface observations, often underused in mesoscale convective research until recently, are the major source of detailed boundary layer data for resolving temperature, moisture, and wind fields. The usefulness of additional surface data was demonstrated with Synoptic Index forecasts, where increased surface data resolution provided an improvement in forecasts of mesoscale convective complexes (Strong et al, 1984). Any increase in upper air resolution should be backed up by an improved surface network. A general rule might be to have surface data sites of about twice the density of the finest upper air resolution. The SESAME data set fail in this regard, especially in the vicinity of the mesoscale upper air network. A denser surface network in quadrant from the mesoscale upper air network, for example, might have resolved source regions for the pre-storm dynamics through surface divergence computations.
6. Mesoscale network planners should consider the problems of merging different data bases before an operation is attempted, since the data methods used can have an impact on the network design itself. This could include the merging of rawinsonde, Doppler wind profiler, and aircraft wind data, or rawinsonde and satellite-derived thermodynamic data.

Many diagnostic studies can be recommended for scale interaction studies of convective storms. Several important ones which arise from the current study follow.

7. Vertical velocity is one of the most important meteorological variables, but it is also one of the most difficult to measure directly. At the one extreme, values of vertical velocity, estimated within severe thunderstorm updrafts to exceed 50 m s^{-1} , are far too violent to sample with penetrating aircraft or other direct means. Fortunately, remote sensing Doppler radar is now alleviating this problem. At the other extreme, the normally small values of synoptic scale and mesoscale environmental values of $1\text{-}10 \text{ cm s}^{-1}$ are several orders of magnitude too low to measure directly with rawinsondes or other current technology. Nevertheless, it is vertical motion of this range of values which results in our major cloud systems. As demonstrated in this study, it is also this range of values which creates the mesoscale environment conditions necessary for a severe convective storm.

Vertical motions of less than 1 m s^{-1} (100 cm s^{-1}) also cannot be resolved with current Doppler radars. The problems of computing vertical motion indirectly, through the wind and thermodynamic fields, were addressed in Chapter 4. It is apparent that additional comparative mesoscale tests of the various means of computing this important quantity are necessary, while also investigating new methods of computation or measurement.

8. No attempt was made to compute the orographic component of pre-storm subsidence and ascent. This study simply concentrated on the total non-storm, environmental motions. Orographic components of vertical motion have been computed in other studies however, and they should be included in future diagnostic work.

9. Scale interactions are essentially energy interactions. Further diagnostic studies of energetics at both the synoptic and mesoscales, as conducted by the Kung and Fuelberg groups, should examine the important pre-storm period in detail, using the multi-scale data suggested above. In fact, it is important to stratify the energetics according to the pre-storm, storm formation, mature storm, and decaying storm periods, with similar and concurrent synoptic stratifications for the cyclones involved over longer periods. Such stratifications would assist in the testing of the additional scale interaction concept presented in Section 8.3.
10. Further time- and space-resolved kinetic energy studies (as opposed to the area-average studies of Kung and Fuelberg) at various scales are necessary to resolve the cause and effect questions posed by downscale and feedback concepts.
11. Surface divergence computations should be included in operational routines used in forecasting convection. This could be a valuable and inexpensive tool for short-term (1-6 hour) forecasts of pre-storm dynamics in the absence of a mesoscale upper air network, since hourly coarse mesoscale (130-km resolution) surface data is always available. Organized surface convergence and low-level ascent was present at least 4-6 hours prior to first radar echoes in the AVE-V case studied here.

12. No mesoscale experiment should be considered complete without a modeling effort. Improvements in mesoscale prediction models can be made through sensitivity tests, and by using real mesoscale data for initialization. The relative importance of local topographic forcing and insolation in various situations can be resolved through modeling tests. Also, mesoscale models are capable of testing the validity of the theories postulated in this thesis.

Regardless of any shortcomings in analysis, this dissertation has highlighted, above all, the immensity of the problem of fully understanding convective storms. In short, it is a scale interaction problem, which can never be fully resolved by focussing theoretical, modeling, or field research efforts only at a small portion of the scale spectrum of atmospheric motions. Remaining significant mesoscale questions can be answered through efforts which include some focus beyond just the cloud scale or mesoscale domains. One should not be satisfied that meteorology has reached the limits of its understanding or predictability of deterministic mesoscale processes. The outlook still looks bright for vast improvements in our ability to understand and forecast the mesoscale.

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APPENDIX I

Planetary and Synoptic Scale Patterns
Associated with Severe Convective Storms

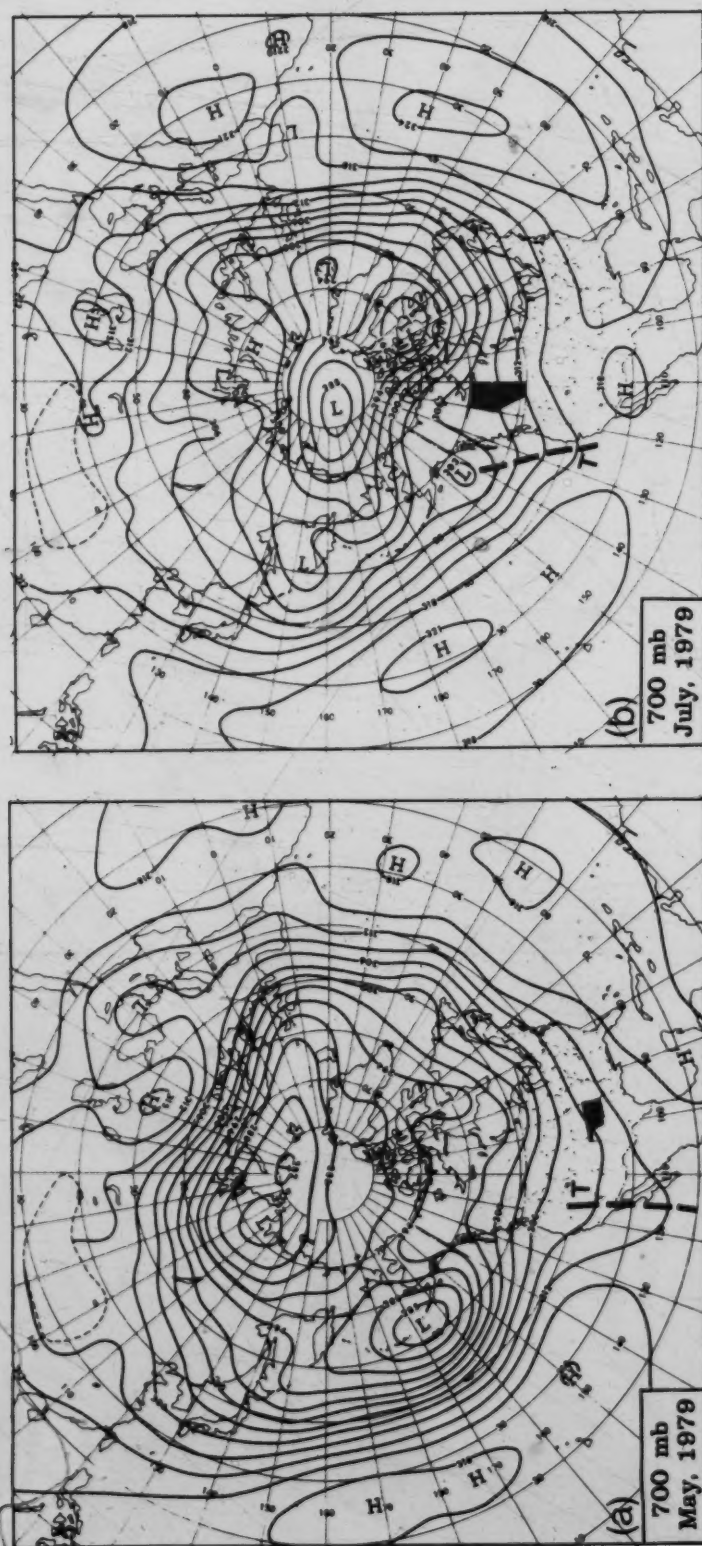


FIGURE A1.1: Mean monthly 700 mb contours (dam) during the peak severe convective weather months for (a) OKLAHOMA - May, 1979 (reproduced from Dickson, 1979); (b) ALBERTA - July, 1979 (reproduced from Wagner, 1979).

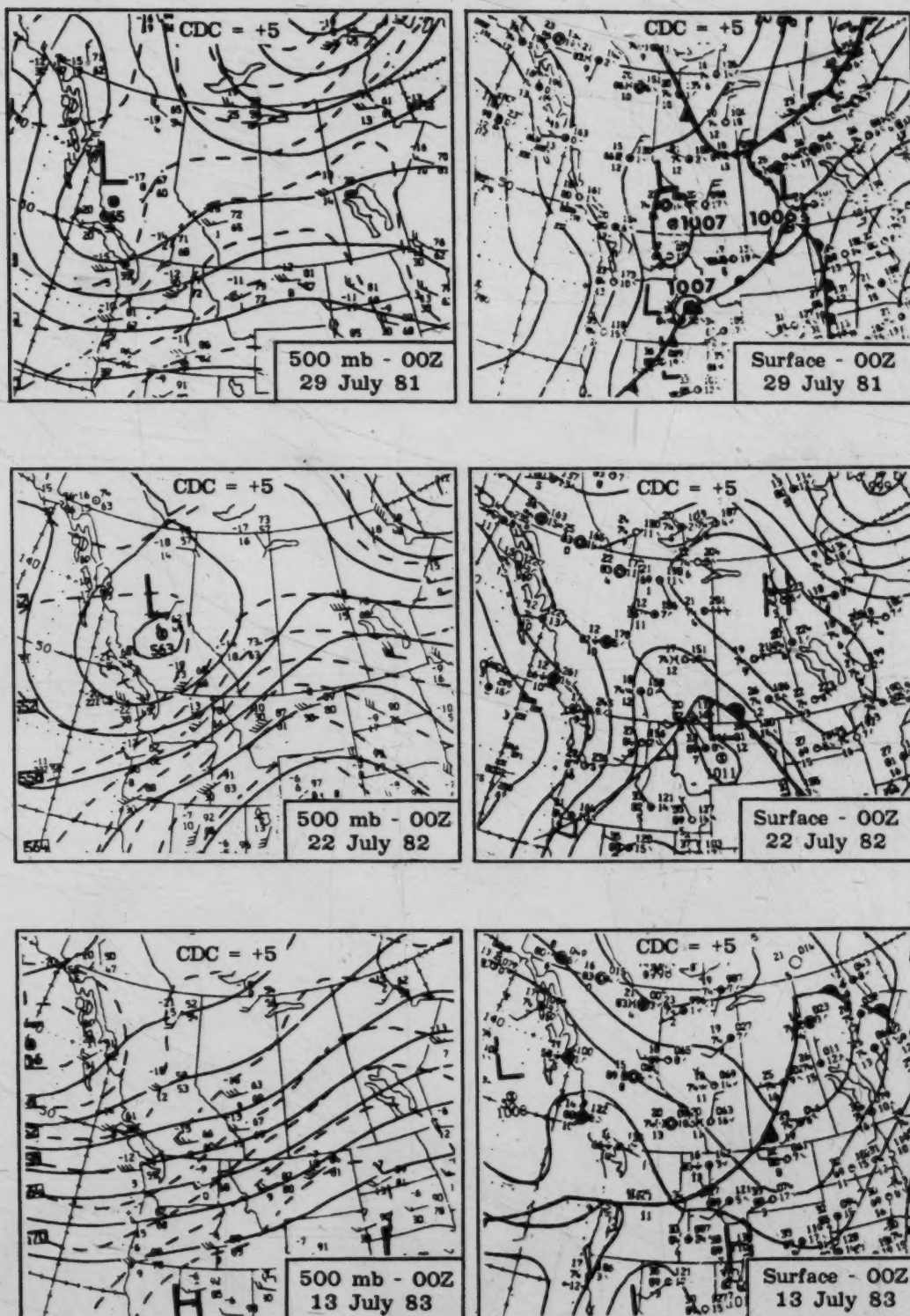


FIGURE A1.2: - caption on next page -

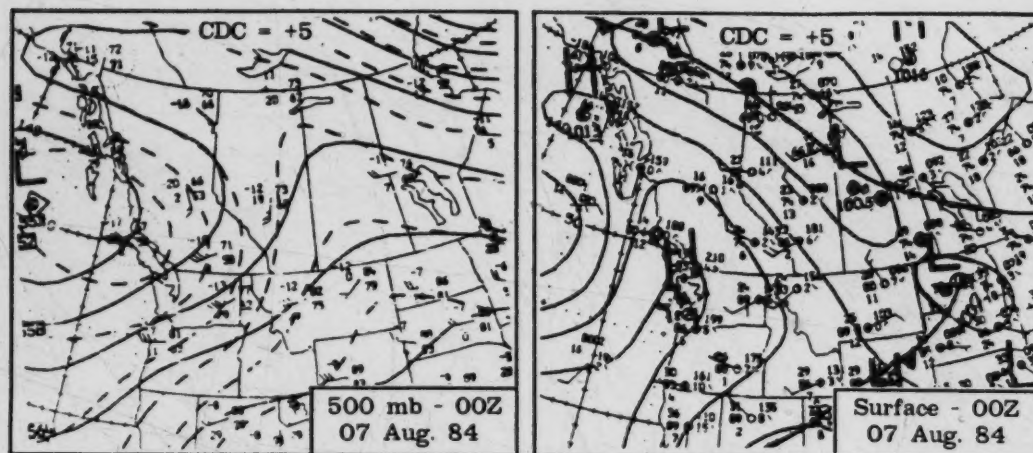


FIGURE A1.2: 500 mb and surface synoptic patterns for the most severe central Alberta hailstorms for each year of 1981-84.

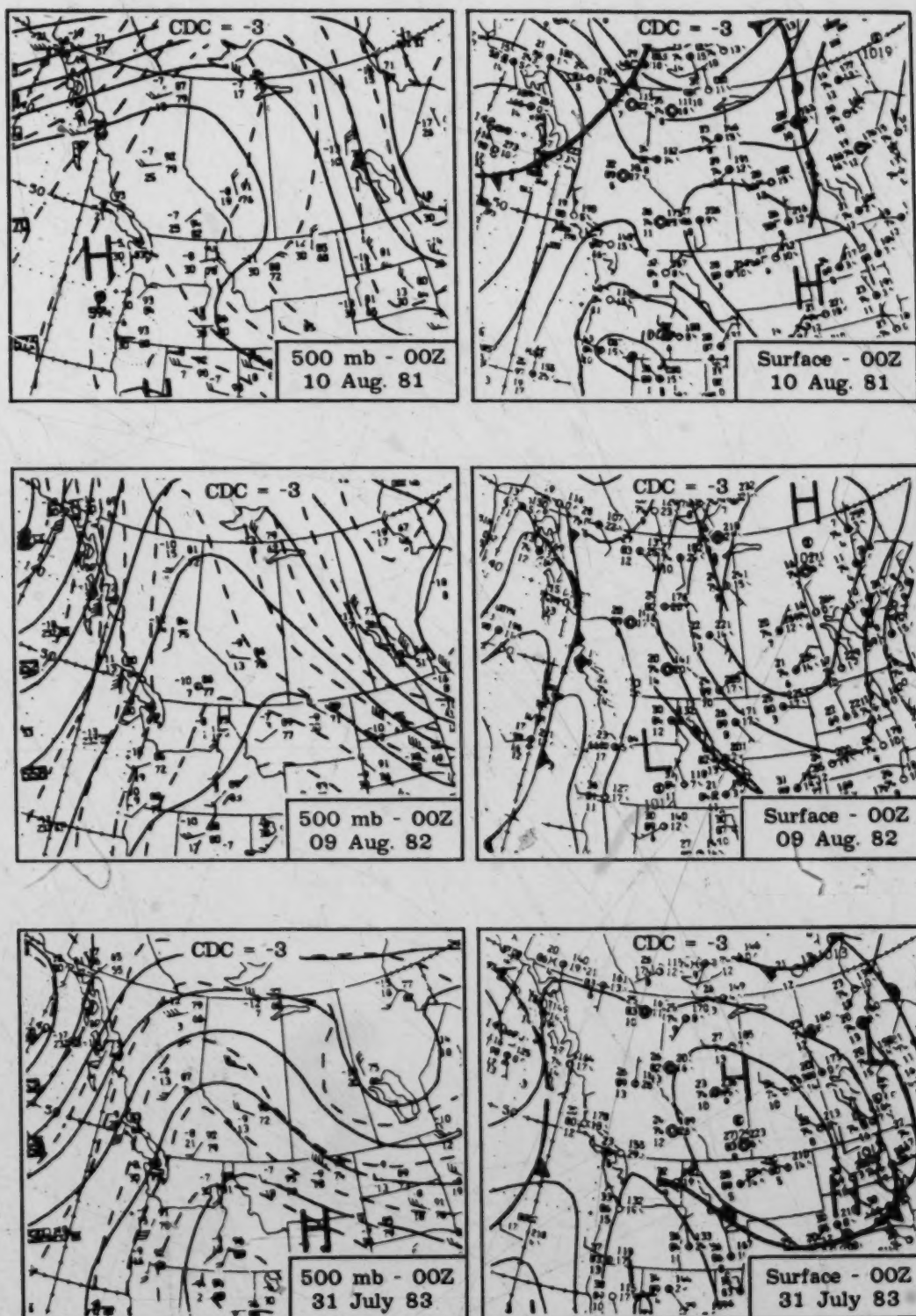


FIGURE A1.3: - caption on next page -

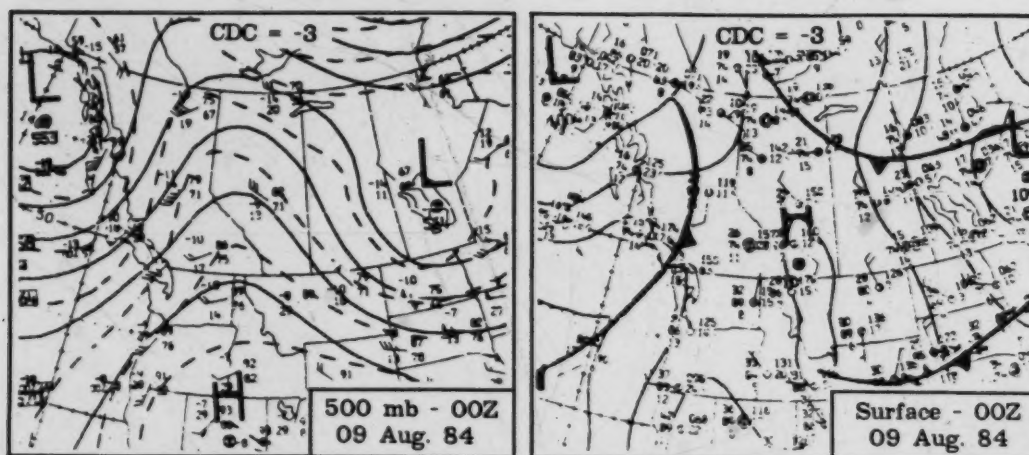


FIGURE A1.3: 500 mb and surface synoptic patterns for the most (convectively) stable days in central Alberta for each year of 1981-84.

APPENDIX II

Data Source Summary
and
Upper Air Station Information

TABLE A2.1: Summary of data types, resolution, and sources for Oklahoma-Texas and Alberta case studies of this research.

Data Type	Region(s)	Approx. Data Spatial (km)	Resolution Temporal (hours)	Data Source
LIMEX-80 upper air	Alberta	(too few sites)	3	Alta. Res. Council, Red Deer, Alberta
all Synoptic scale anal.	Canada, U.S.	-130 {sfc} 4-500 {u/a}	6 12	SDSD, NOAA, Washington
GOES-West IR sat. data	Alberta	8 {N-S} 16 {W-E}	0.5 0.5	SDSD, NOAA, Washington
NOAA-6 sat. images	Alberta	1-2	3-12	Univ. of Alta. and AES, Edmonton
SESAME upper air	central U.S.	370 {synop.} 80 {meso}	3 3	NASA, Marshall Cntr., Huntsville, Al.
SESAME surface data	North America	130	1	Florida State Univ. Tallahassee, Fl.
GOES-East Vsb/IR images	eastern U.S.	2-4	1	SDSD, NOAA, Washington
WSR-57 Radar PPIs	Oklahoma	0.5-2	1.5 mins.	NSSL, Norman, Okla.
Radar Composites	central U.S.	unknown	1-2	NASA, Marshall Cntr., Huntsville, Al.

TABLE A2.2: Synoptic and mesoscale upper air station information, including 3-letter station identifier, latitude and longitude (degrees), elevation (meters MSL), station number, and location in alphabetical order by state and town.

SYNOPTIC Stations:

sID	Lat.	Long.	Elev.	No.	Location
CKL	32.90	87.25	141.	229	Centerville, Alabama
FLG	35.13	111.67	835.	375	Flagstaff, Arizona
TUS	32.12	110.93	802.	274	Tucson, Arizona
LIT	34.73	92.23	78.	340	Little Rock, Arkansas
DEN	39.75	104.86	1625.	469	Denver, Colorado
GJT	39.11	108.53	1480.	476	Grand Junction, Colorado
BOI	43.57	116.22	871.	681	Boise, Idaho
PIA	40.66	89.68	201.	532	Peoria, Illinois
SLO	38.65	88.97	171.	433	Salem, Illinois
DBQ	42.40	90.70	327.	547	Dubuque, Iowa
DDC	37.76	99.96	791.	451	Dodge City, Kansas
TOP	39.06	95.63	268.	456	Topeka, Kansas
BVE	29.33	89.39	1.	232	Boothville, Louisiana
LCH	30.11	93.21	4.	240	Lake Charles, Louisiana
FNT	42.97	83.75	233.	637	Flint, Michigan
SSM	46.47	84.37	221.	734	Sault Ste. Marie, Michigan
INL	48.57	93.38	361.	747	International Falls, Minnesota
MSP	44.88	93.21	256.	658	Minneapolis, Minnesota
STC	45.55	94.07	312.	655	St. Cloud, Minnesota
JAN	32.31	90.08	94.	235	Jackson, Mississippi
UMN	36.88	93.90	406.	349	Monett, Missouri
GGW	48.22	106.62	700.	768	Glasgow, Montana
GTF	47.48	111.39	1115.	775	Great Falls, Montana
LBF	41.13	100.68	847.	562	North Platte, Nebraska
OMA	41.30	95.90	299.	553	Omaha, Nebraska
ABQ	35.05	106.61	1631.	365	Albuquerque, New Mexico
BUF	42.93	78.73	215.	528	Buffalo, New York
BIS	46.77	100.75	506.	764	Bismark, North Dakota
DAY	39.90	84.20	306.	429	Dayton, Ohio
OKC	35.40	97.60	391.	353	Oklahoma City, Oklahoma
PIT	40.50	80.22	373.	520	Pittsburgh, Pennsylvania
HON	44.38	98.22	393.	654	Huron, South Dakota
RAP	44.05	103.07	966.	662	Rapid City A., South Dakota
BNA	36.11	86.68	182.	327	Nashville, Tennessee
AMA	35.23	101.70	1099.	363	Amarillo, Texas
DRT	29.36	100.91	310.	261	Del Rio, Texas
ELP	31.80	106.39	1206.	270	El Paso, Texas
GGG	32.38	94.71	111.	244	Longview, Texas
MAF	31.95	102.18	874.	265	Midland, Texas
VCT	28.85	96.91	32.	255	Victoria, Texas
SEP	32.22	98.18	398.	260	Stephenville, Texas

SLC	40.77	111.97	1288.	572	Salt Lake City, Utah
HTS	38.37	82.55	255.	425	Huntington, West Virginia
GRB	44.48	88.13	212.	645	Green Bay, Wisconsin
LND	42.82	108.73	1694.	576	Lander, Wyoming
CUU	28.70	105.96	1354.	225	Chihuahua, Mexico

SESAME Mesoscale Stations, May-June, 1979:

sID	Lat.	Long.	Elev.	No.	Location
ADA	34.80	96.67	311.	020	Ada, Oklahoma
LTS	34.64	99.26	420.	021	Altus, Oklahoma
CAN	35.90	100.40	735.	022	Canadian, Texas
CHE	35.61	99.70	622.	023	Cheyenne, Oklahoma
CHK	35.10	97.96	352.	024	Chickasha, Oklahoma
CDS	34.43	100.28	595.	025	Childress, Texas
CSM	35.36	99.20	584.	026	Clinton Sherman, Oklahoma
EMC	34.65	97.41	320.	027	Elmore City, Oklahoma
FSI	34.64	98.39	360.	028	Fort Sill, Oklahoma
GAG	36.30	99.76	678.	029	Gage, Oklahoma
HEA	34.25	97.47	291.	030	Healdton, Oklahoma
HEN	36.05	97.89	343.	031	Hennessey, Oklahoma
HNT	35.47	98.40	507.	032	Hinton, Oklahoma
TVY	35.57	97.49	363.	033	KTVY, Oklahoma City, Oklahoma
MTV	35.10	98.72	416.	034	Mountain View, Oklahoma
SEL	36.10	98.92	588.	036	Seiling, Oklahoma
SHM	35.24	100.19	721.	037	Shamrock, Texas
SUD	35.78	96.66	272.	038	Stroud, Oklahoma
SPS	33.96	98.48	309.	039	Wichita Falls, Texas

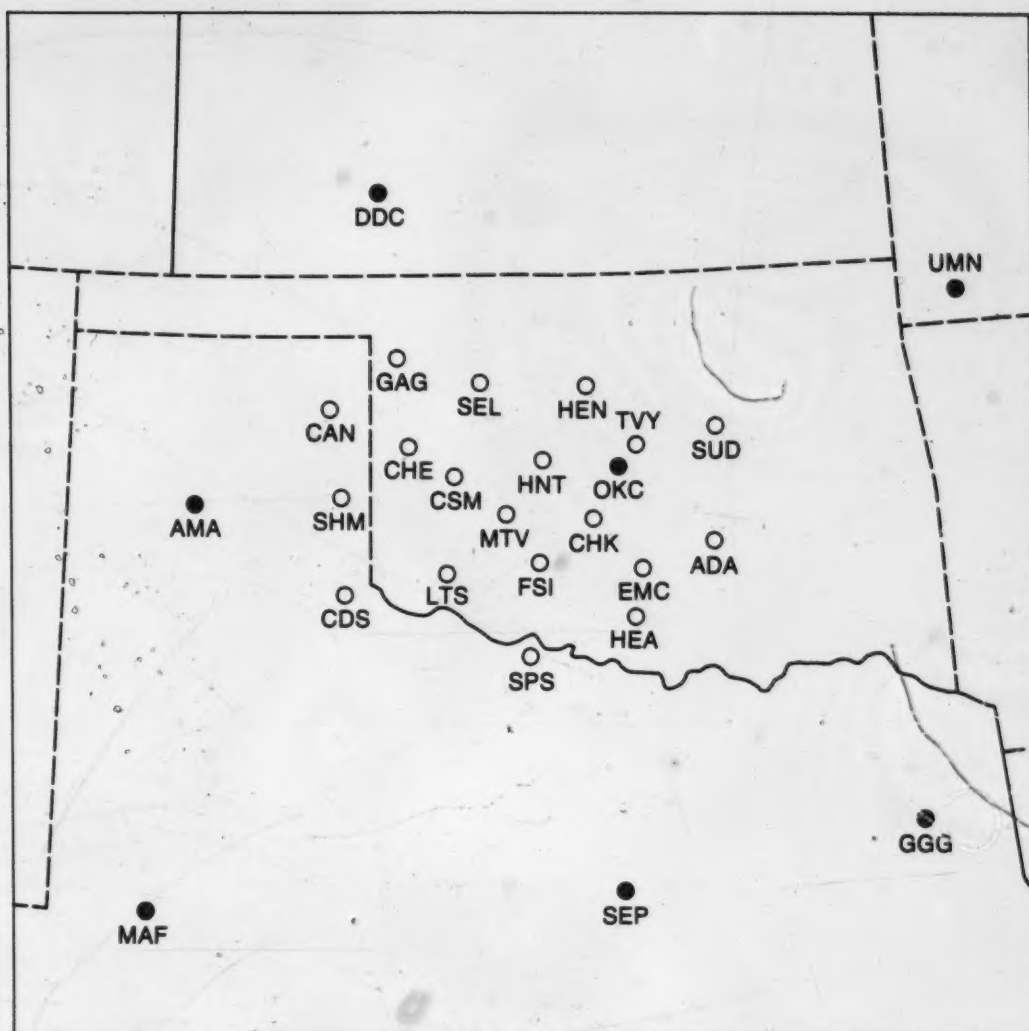


FIGURE A2.1: SESAME-AVE-IV/V/VI mesoscale upper air network.

APPENDIX III



Development and Operational Use of
the Synoptic Index of Convection (Sc4)

Development and Operational Use of the Synoptic Index of Convection (Sc4)

The formulation of the Sc4 and its forerunner, the Sc10, have been described elsewhere (Strong, 1979; Strong and Wilson, 1981, 1983a), but are summarized here for ease of reference.

A3.1 DERIVATION OF THE Sc4

The Synoptic Index is currently computed from four predictor variables, each independently correlated to a predictand called Convective Day Category (CDC), as defined in Table 4.4. The CDC was designed as a single number which could roughly summarize the maximum degree of convective intensity occurring for a day over the ARC Hail Project operations area (AHP). Thus, it is important to realize that a point value of CDC (likewise, of the predictor, Sc4) represents conditions for "130 km radius from that point", and ideally represents the "maximum" convective intensity "for the whole day" (actually 1800Z to 1200Z following morning).

The simplest way to describe this convective intensity with a single value was to use maximum hail size reported within AHP, since many years of records for this variable are available with spatial resolution of 1-5 km. Hail size is related to storm updraft strength, liquid water content of the cloud, storm duration, and so on. For no-hail days, hourly and special surface observations are used to determine the areal coverage of showers and/or deep convective clouds. The categories were chosen so that the distribution of CDC for a season approached a Poisson type of distribution (Strong, 1979). Since virtually all possible predictor variables would be similarly distributed, the best possible statistical correlations with CDC could thereby be obtained.

Almost 100 predictor variables were tested for statistical correlation to CDC. They were selected on sound physical bases; i.e., some known relation to instability. The Stability Tendency Equation (see Iribarne and Godson, 1973 for derivation), for example, provides a host of possible predictors; it can be stated as:

$$\begin{aligned} \frac{\partial \gamma}{\partial t} = & - \frac{\partial}{\partial z} \left(\frac{1}{c_p} \frac{dQ}{dt} \right) + \frac{\partial}{\partial z} \left[u \frac{\partial T}{\partial x} + v \frac{\partial T}{\partial y} \right] \\ & + \left(\Gamma_d - \gamma \right) \frac{\partial w}{\partial z} - \omega \frac{\partial \gamma}{\partial z} \end{aligned} \quad (A3.1)$$

The original Synoptic Index, the Sc10, was then a linear combination of 10 of the highest correlating variables (with CDC) for which objective forecasts could also be made (Strong, 1979). Only one value of the Sc10 was computed for all of the AHP operations area

(130-km radius from Red Deer, Alberta). This version was used successfully for AHP operations from 1978-82.

During this time, it was found that an index composed of fewer and less cumbersome variables (4-7), provided forecasts which were almost as reliable as the Sc10 (Strong and Wilson, 1981). One of these, the Sc4, was comprised of two synoptic and two instability variables. Unlike the Sc10, which required input from a radiosonde sounding, the Sc4 instability variables used any surface data and a simple forecast of 500 mb temperature. Thus, it is easily computed, so that a regional version of the Synoptic Index, providing forecast maps of maximum convective intensity became possible (Strong and Wilson, 1983a,b). Shortly thereafter, forecasts of maximum (negative) advection of Sc4 extended its capability to predictions of storm initiation location, motion, relative convective complex size, and the location of maximum intensity at storm maturity (Strong and Wilson, 1983c). At the same time, inclusion of surface moisture advection and judicious use of satellite/radar data, extended its usefulness down to the mesoscale time and space scales (Strong et al, 1984). Two examples of these capabilities were shown in Section 1.3.8.

The predictor equation for the Sc4 simply sums the contributions of the predictor variables according to

$$P = \sum_{i=1}^n w_i (a_i p_i + b_i) \quad (A3.2)$$

where $n = 10$ for the Sc10, $= 4$ for the Sc4. The a_i, b_i are the least-squares coefficients of approximation of p_i to CDC; w_i represents the weighting coefficients, each determined from the relative degree of correlation by

$$w_i = \frac{|\rho_i|}{\sum_{k=1}^n |\rho_k|} \quad (A3.3)$$

where the ρ_i are linear correlation coefficients and $\sum w_i = 1.0$

A final transformation of the predictor Equation (A3.2) is performed to yield

$$Sc_4 = 3.73 P - 1.38 \quad (A3.4)$$

which forces the relation into a one-to-one correspondence with CDC.

The Sc4 is not computed if more than one predictor variable is missing. In the event that one variable is missing, the other three are re-weighted in the computations using Equation (A3.3).

The four predictor variables were chosen on the basis of high correlation with CDC, a desirability to blend synoptic scale variables with sub-synoptic scale instability parameters, and ease of automation. The two instability variables are the EGDEX (Checkwirth, 1971) and the Surface Parcel Lifted Index (SPLI). The synoptic variables are 24-hour changes in the 1000-500mb thickness, and in the 700mb temperature. The changes are computed from 1200Z current day to 1200Z the following day. The EGDEX computes the difference between the wet bulb potential temperature (Theta-W) of the surface parcel and the 2400Z 500mb Theta-W (assuming saturation). There is also a 15-2100Z diurnal adjustment applied to the surface Theta-W for each station, based on climatological statistics by month and wind direction. The SPLI is the difference between the 2400Z 500mb temperature and the surface parcel temperature at 500mb. Positive values of EGDEX or SPLI represent instability, negative values stability. Both use the same surface and 500mb data, and can be frequently updated using hourly surface data.

Table 4.4 provided definitions of the nine discrete CDC categories. Table A3.1 shows the most recent correlation and weighting statistics for the predictor variables used in current Sc4 computations (Strong and Wilson, 1983a), and compares overall correlations of the Sc4 and Sc10 with CDC over central Alberta.

TABLE A3.1: Correlation and weighting statistics for Sc4 predictor variables with CDC.

Predictor Variable	Correlation With CDC	Weighting Coefficients
EGDEX	0.577	0.288
24-hr. 700mb Temperature Change (12-1200Z)	-0.492	0.246
24-hr. 1000-500mb Thickness Change (12-1200Z)	-0.479	0.239
Surface Parcel Lifted Index	0.456	0.227

A3.2 GUIDELINES FOR FORECASTING WITH THE SYNOPTIC INDEX

Following several years of experience in using the "regional version" of the Synoptic Index (Sc4), a number of guidelines as to its interpretation and use can be made. Reference is made to Figures 1.7 and 1.8 as examples in following through these guidelines. While the following nine rules will not always guarantee an accurate forecast,

familiarity with the Sc4, including its limitations due to data resolution, have prevented extremes in poor forecasts during its three years of service to date.

A3.2.1 General Threat Areas and Intensity

1. Use the +3 SC4 contour to delineate severe weather threat areas.
Note: CDC = +3 implies walnut size hail "within a 130-km radius".

2. Forecast centres of SC4 maximum, or lines of SC4 maximum, are, to sub-synoptic scale resolution, where storms should attain their maximum intensity.

Note : the Sc4 spatial resolution is only of sub-synoptic scale because:

- a) the CDC (Convective Day Category) is defined in this way, and;
- b) the finest data resolution (to date) has been the sub-synoptic scale surface network - resolution of predictor variables which go into the SC4 are as follows : 1. EGDEX - 130 km (surface); 2. SPLI - 130 km (surface); 3. 24-hour change in 700 mb temperature - 500 km (upper air); and 4. 24-hour change in 1000-500 mb thickness - 500 km (upper air).

The RAINSAT project (Strong et al., 1984) showed that improved surface (and Sc4) resolution could be obtained over the foothills of the Alberta Rockies, by using Forestry weather reports, having an average spacing of about 30 km; however, these are only reported twice daily (at 0730 and 2030 MDT).

A3.2.2 Storm Initiation

3. One can forecast the region of Sub-synoptic Scale Convective Complex (SSCC) initiation as the region of negative SC4 advection (strong Sc4 gradient upwind from the Sc4 maximum). The 700 mb flow (close to cloud base) is used to determine the advection.

4. Fields of "either" surface mixing ratio¹ OR Theta-E also provide an estimate of where storms are likely to initiate, as well as short-term motion. Mixing ratio and Theta-E tend to be a maximum on the inflow side of storm complexes (generally the east side, or downwind). Therefore, the region slightly upwind of the moisture maximum (or line of maximum moisture) has a high potential for storm formation. Low-level winds (surface, or 850 mb if in mountains) are used in this case to determine advection. The

¹ Surface mixing ratio is a more useful measure for the pre-storm (morning forecast) period, if only because Theta-E changes quite rapidly with early heating. Also, mixing ratio provides areal estimates of representative moisture to apply to a sounding analysis, and can be updated hourly. Once a storm has formed, experience shows that Theta-E is more useful, perhaps because it is a direct measure of thermodynamic potential for storms now formed.

region within 5C from the "ridge of" Theta-E maximum (Figures 1.8d, 6.25b, and 6.26b) appears to be most favourable.

5. To combine the forecast SC4 and current Theta-E advection fields, use the intersection of these regions of strongest advection to further delineate the forecast region of storm initiation. Of course, lines of maximum convergence in the surface wind field will also help to delineate regions of formation.

A3.2.3 Storm Complex Propagation

6. Once formed, individual storms will tend to move towards the line of Theta-E maximum, at least within sub-synoptic scale resolution or about 65 km. The convective complex (SSCC, or cluster of storms) will tend to propagate towards the (line of) SC4 maximum, and to attain its maximum intensity there.
7. The most severe storm of any large complex or line will tend to be the one moving towards the actual center of maximum SC4.

A3.2.4 Fine-tuning the Forecast Using Satellite Imagery and Radar

8. If GOES imagery and radar data are available, the earlier forecasts can be fine-tuned once initial convection has occurred. Cumulus Congestus (TCU) clouds are often observed 1/4-1 hour prior to first radar echoes (personal experience). The larger (synoptic to sub-synoptic) scale organization of this TCU is usually evident from satellite imagery, so that this scale of 'complex region' can often be identified within half an hour of first occurrence. Once clouds reach radar detectability, individual mesoscale organizations can be quickly identified. From these two stages, updated trend forecasts can be made, using the latest Theta-E field, and any updated SC4 field; i.e.; storms tend to move towards the (line of) maximum Theta-E.

A3.2.5 Storm Decay

9. Storms tend to weaken once past the line of maximum SC4, or sooner if the supply of moisture (as indicated by the field of surface mixing ratio or Theta-E) is exhausted by cycling through the storm system.

APPENDIX IV

Supporting Sub-synoptic to Mesoscale Analyses

Appendix IVa: Sub-synoptic scale Surface Analyses for 1200 UTC, 20 May, 1979, through 1100 UTC, 21 May, 1979.

Appendix IVb: Mesoscale 300, 500, 700, and 850 mb isotach and jet streak analyses for 1100-1100 UTC, 20-21 May, 1979.

Appendix IVc: Mesoscale 12- and 24-hour changes in wind speed (m s^{-1}) at 300, 500, 700, and 850 mb, for 1100-2300, 2300-1100, and 1100-1100 UTC, 20-21 May, 1979.

APPENDIX IVa - Hourly Surface Analyses, 20-21 May, 1979

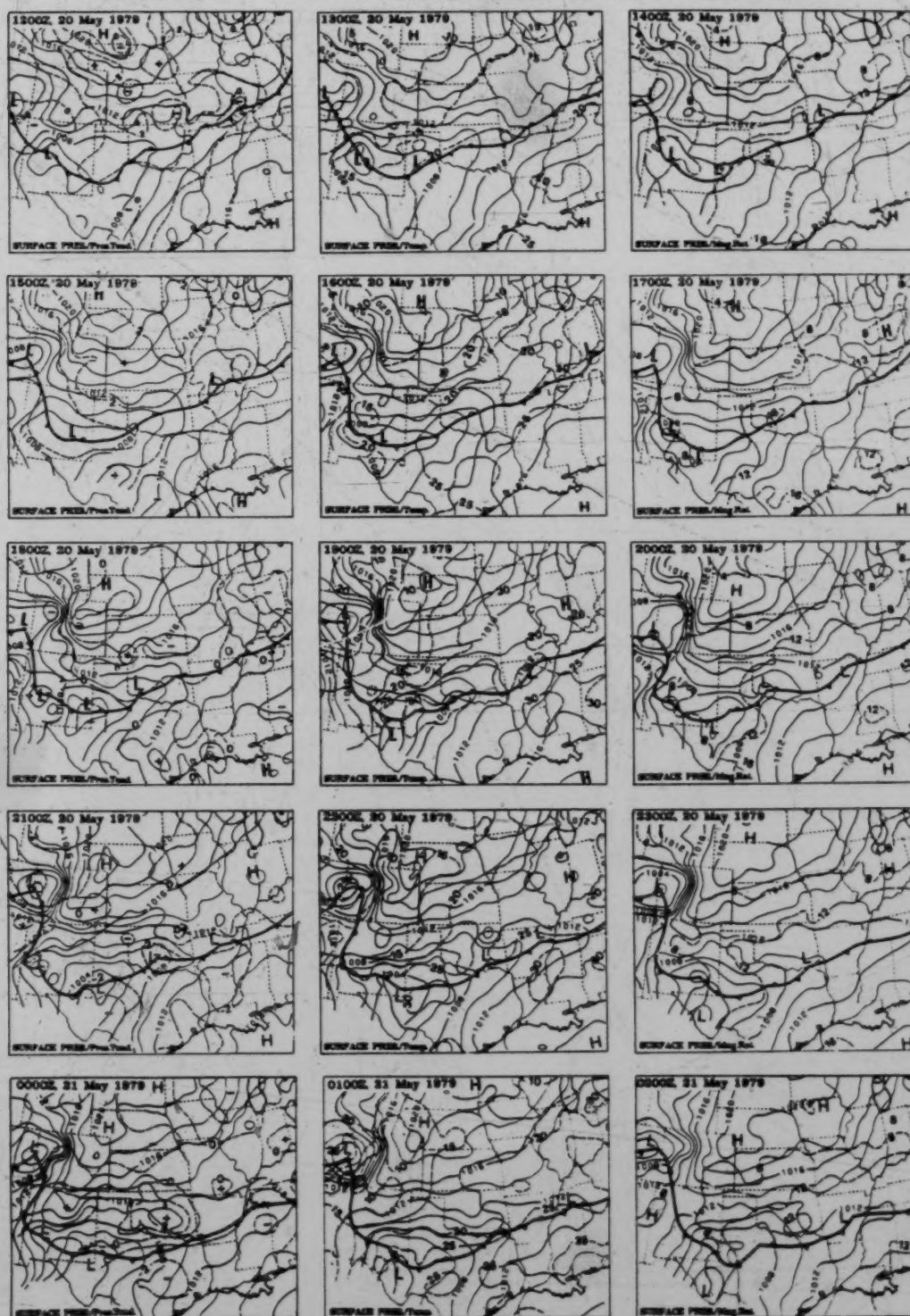


Figure A4.1: (caption next page)

APPENDIX IVa - (continued)

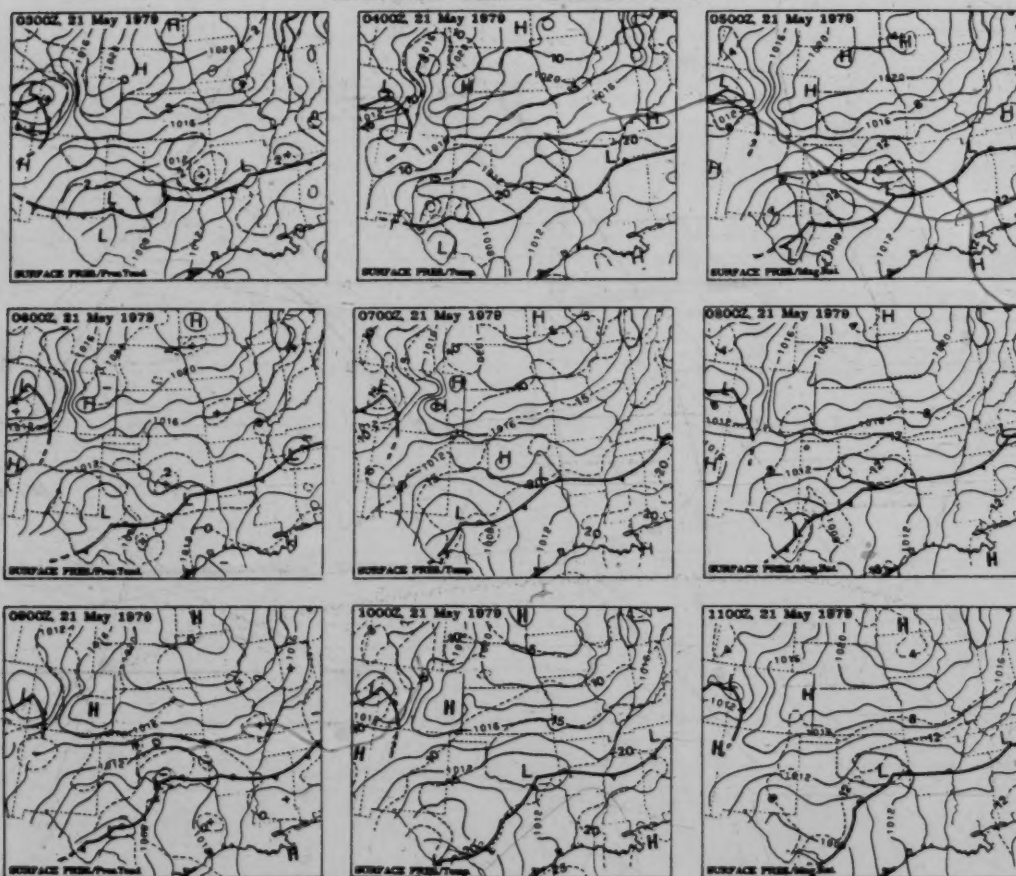


FIGURE A4.1: Hourly surface analyses for 1200 UTC, 20 May, 1979, through 1100 UTC, 21 May, 1979. These include surface front positions, isobars at 2-mb intervals, and, at alternate times, (broken) contours of pressure tendency ($\text{mb}/3 \text{ hours}$), temperature ($^{\circ}\text{C}$), and mixing ratio (g kg^{-1}), commencing at 1200, 1300, and 1400 UTC respectively.

Appendix IVb

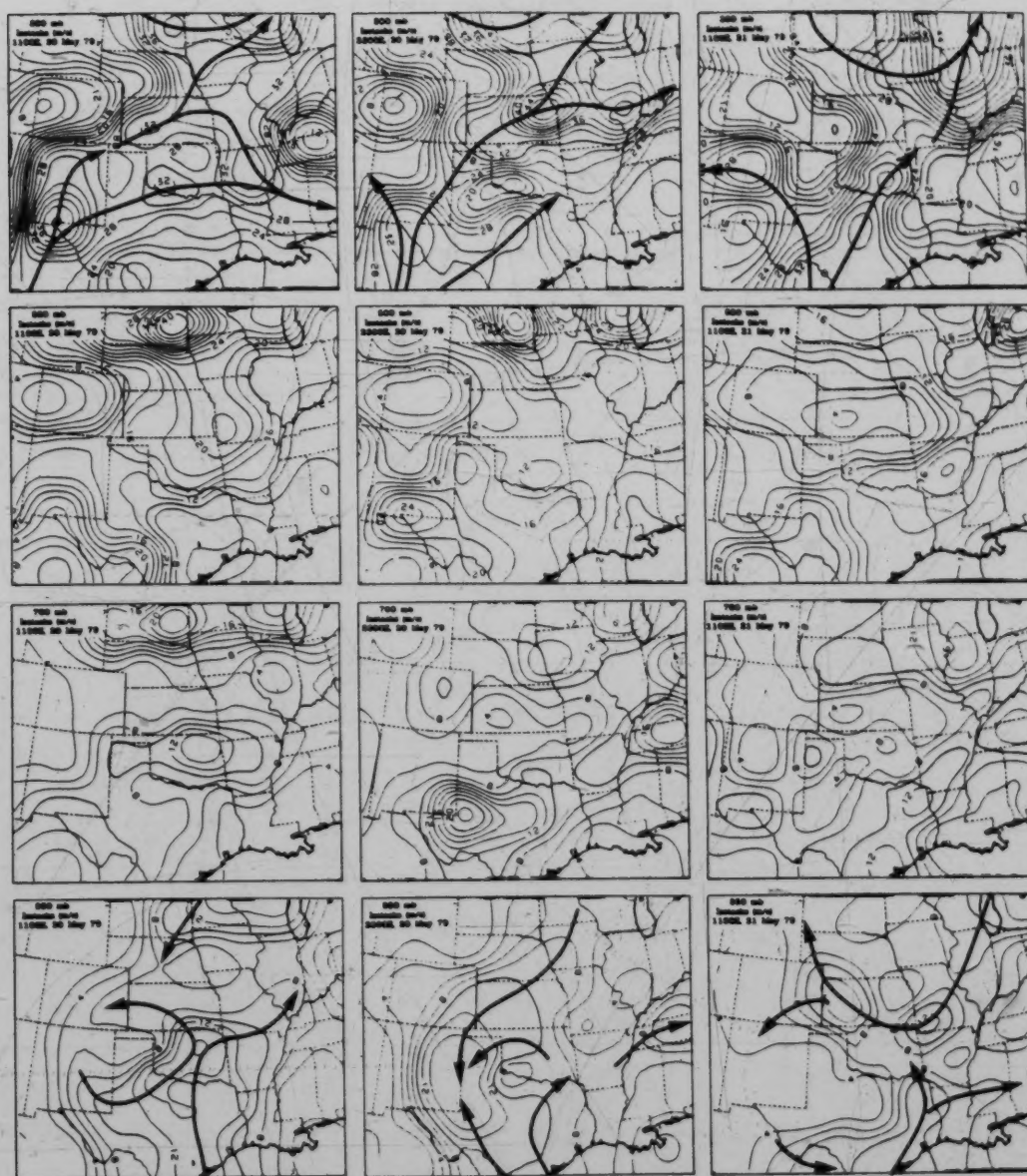


FIGURE A4.2: Mesoscale 300, 500, 700, and 850 mb isotach (m s^{-1}) and jet streak analyses for 1100-1100 UTC, 20-21 May, 1979.

Appendix IVc

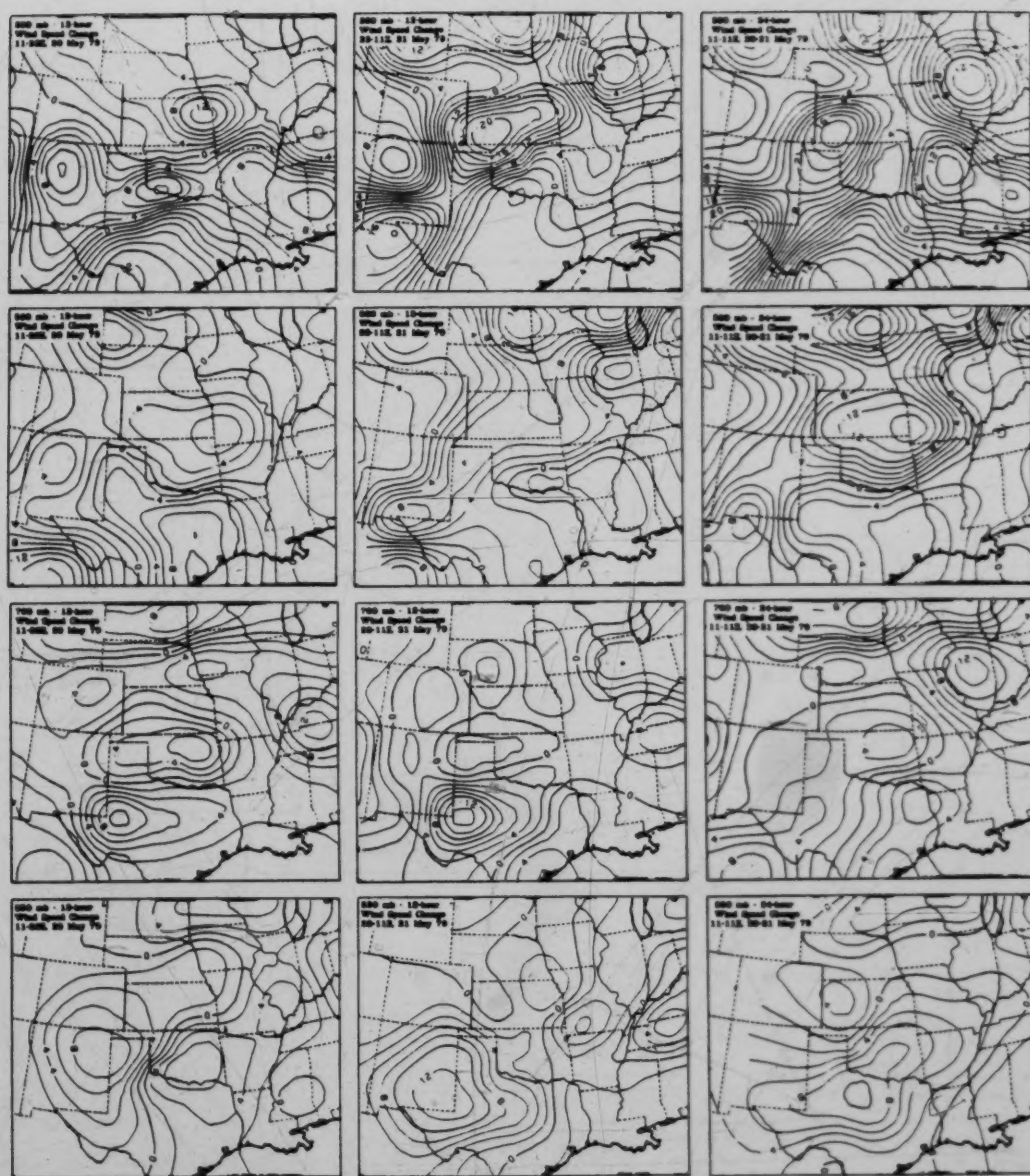


FIGURE A4.3: Mesoscale 12- and 24-hour changes in wind speed (m s^{-1}) at 300, 500, 700, and 850 mb, for 1100-2300, 2300-1100, and 1100-1100 UTC, 20-21 May, 1979.

APPENDIX V

Quantitative Comparison (Inter-correlations) of Sounding Parameters

This table includes all data from the synoptic as well as mesoscale network stations of this study. Separate statistics are provided for the 1400 and 1700 sounding sets. The top right half of the correlation matrix contains the correlations for 1400 UTC, the bottom left half is for 1700 UTC.

Means and Standard Deviations:

	Tc 1	Avg.MR 2	P(LFC) 3	T(LFC) 4	Theta-W 5	LI(500) 6	PcIt 7	TcIt 8	P(sfc) 9	T(sfc) 10	Ps-Pci 11	Tc-Ts 12	Tct-Tcb 13
1400Z													
MEAN:	32.83	12.84	760.85	13.36	23.12	6.81	825.60	18.02	955.93	21.72	128.20	11.11	17.01
S.D.:	2.65	1.95	35.78	2.88	1.56	2.13	31.90	3.02	24.85	3.46	35.99	2.84	3.24
1700Z													
MEAN:	32.06	13.14	772.12	13.91	23.18	6.96	786.80	15.30	955.96	25.05	167.00	6.98	14.87
S.D.:	3.19	1.80	44.61	2.83	1.43	2.02	42.52	3.71	25.65	4.30	41.18	3.64	3.47

CORRELATION MATRIX:

	Tc 1	Avg.MR 2	P(LFC) 3	T(LFC) 4	Theta-W 5	LI(500) 6	PcIt 7	TcIt 8	P(sfc) 9	T(sfc) 10	Ps-Pci 11	Tc-Ts 12	Tct-Tcb 13
1	1	0.590	0.002	0.520	0.824	0.574	0.216	0.701	0.370	0.598	-0.070	0.205	0.618
2	0.374	1	0.721	0.985	0.832	0.876	0.296	0.572	0.677	0.788	0.221	-0.410	0.636
3	-0.312	0.652	1	0.811	0.253	0.484	0.150	0.051	0.790	0.447	0.512	-0.544	0.198
4	0.248	0.942	0.808	1	0.754	0.831	0.257	0.490	0.742	0.759	0.316	-0.441	0.567
5	0.798	0.764	0.074	0.640	1	0.898	0.350	0.786	0.366	0.791	-0.113	-0.196	0.763
6	0.563	0.807	0.233	0.722	0.927	1	0.317	0.642	0.428	0.814	0.075	-0.458	0.681
7	0.448	0.616	0.181	0.549	0.629	0.577	1	0.734	0.329	0.401	-0.590	-0.301	0.695
8	0.679	0.630	-0.011	0.526	0.855	0.765	0.895	1	0.272	0.727	-0.451	-0.257	0.942
9	0.368	0.721	0.700	0.805	0.428	0.476	0.397	0.305	1	0.564	0.448	-0.343	0.388
10	0.559	0.594	0.272	0.598	0.650	0.689	0.589	0.619	0.597	1	0.054	-0.663	0.831
11	-0.237	-0.375	0.077	-0.242	-0.490	-0.450	-0.758	-0.736	0.110	-0.285	1	-0.132	-0.328
12	0.211	-0.370	-0.591	-0.488	-0.069	-0.235	-0.313	-0.144	-0.382	-0.692	0.135	1	-0.465
13	0.613	0.681	0.141	0.627	0.807	0.761	0.826	0.895	0.434	0.756	-0.579	-0.367	1

V I T A

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